Improving global glacier modelling by the inclusion of parameterised subgrid hypsometry within a three-dimensional, dynamical ice sheet model.

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Chapter 1

Introduction

An o er ie of this study

The study begins by considering how glaciers work conceptually, why improvements in modelling them are of interest, and how such improvements will be attempted within this study. Chapter 2 examines the processes that are involved in glaciers in more depth, previous attempts to model them and their weaknesses, and closes by describing the equations involved in modelling the dynamics and thermodynamics of glaciers. Discussion in chapter 3 focuses on the two main parameterisations used: the degree-day scheme and the subgrid hypsometric scheme. It is the inclusion of this latter parameterisation that led to the choice of model because of the improvements to glacier modelling it can o er. Chapter 4 covers the approaches used

$\mathbf{k}$ is used for the three-dimensional case. Time is denoted by t, temperature by T and (u, v, w) are the components of the three-dimensional velocity field, $v_{\mathbf{k}}(\lambda, \theta, z, t)$, in which λ is the longitude, θ is the co-latitude and z is the elevation. The two-dimensional equivalent is $v_{\mathbf{k}}(\lambda, \theta)$ with components (u, v).

Measures of height or length in this document are metric and wherever a appears as a unit it refers to annum. The earth is split into 360 ° of longitude, and 180 ° of latitude. The terms "minutes" and "arcsecs" are measures of distance, in spherical co-ordinates. Each degree is then split into 60 minutes, and each minute is split into 60 arc-seconds. The co-latitude is the angle between any latitude and the North Pole. London has a latitude of 51 °N, which is a co-latitude of 39 °, and Sydney has a latitude of 34 °S, which is a co-latitude of 124 °.

What is a Glacier

The OED defines a glacier as "A slowly moving mass of ice formed by accumulation and compaction of snow on mountains or near the poles". The health of an individual glacier can be measured by its mass balance; this is the difference between how much ice is gained during the year (accumulation), and how much is lost (ablation). When considering a mountain glacier in balance, as in Figure 1.1, then at high elevations there exists an area of net accumulation, and at low elevations there is an area of net ablation. This is because as altitude increases, the air temperature falls, and hence precipitation is more likely to fall as snow and less likely to melt at the surface, and vice versa as altitude decreases.

To compensate for these net imbalances there is a flux of ice down slope, so that the snow that falls on the top of a glacier today, will compact to form ice (in various stages), then be passed down slope, and eventually melt. wn sludtua6

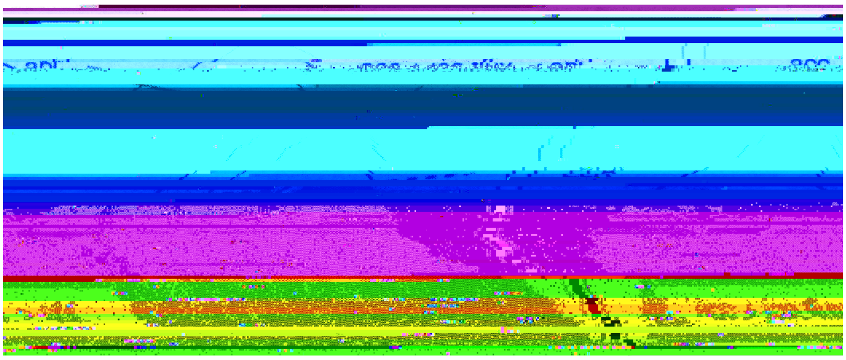

Figure 1.1: from Marshall & Clarke, 1999

little snowfall, with most

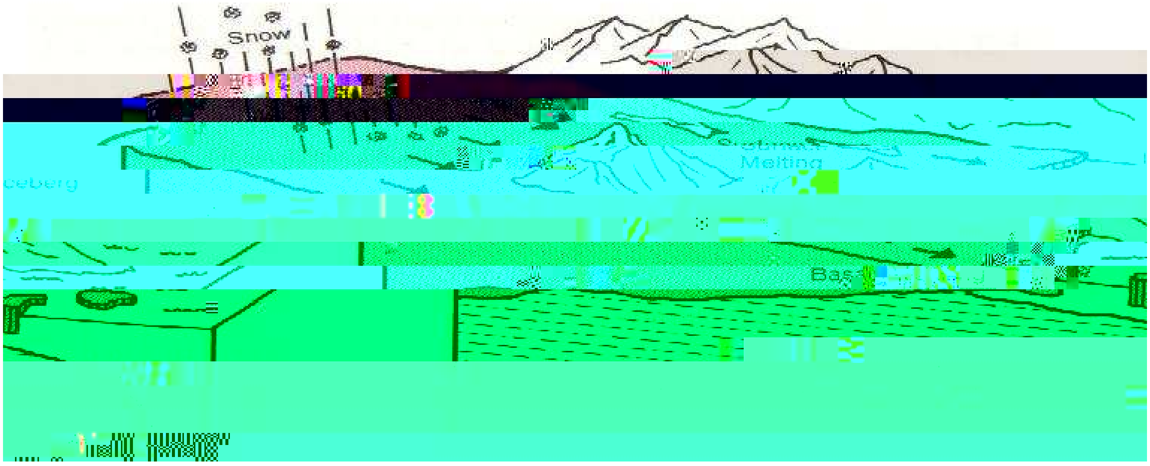

Figure 1.2: from Trenberth, 1999

surrounding sea water, which is why sea-ice floats.

The second form of ice-mass that occurs over the sea is an Ice Shelf, as illustrated in Figure 1.2. It is like an ice sheet that has floated out to sea; it is distinct from sea ice because some part of it remains attached to land - grounded - even though this may be below sea level, such as the Larsen B ice shelf in Antarctica. At the terminus of an ice shelf, as illustrated in Figure 1.2 the ice may disintegrate into floating icebergs. The production of icebergs, known as calving can occur wherever a glacier or ice sheet meets the sea, or where an ice shelf exists. Calving is key in modelling Antarctica because under present climatic conditions it is the only form of mass loss that occurs on this ice sheet.

Ice shelves and sea-ice must both displace sea water to float and so the sea level has already adjusted to carrying their mass. If they should melt sea levels would remain the same, and so we do not need to take direct account of them when modelling sea level rise. The loss of sea ice can feedback on to sea levels by the albedo effect, which is discussed in section 2.1, but this effect is beyond the scope of this study.

Why are we interested in Glaciers

Sea levels affect millions of people; a large portion of the population lives on or near the coast. Therefore improvements in any of the estimates of contributions to sea level rise are of interest. With the increased temperatures and change in water and land use associated with climate

In Antarctica's case the flux of ice is low compared to its total volume, so it takes a long time for changes to filter through. Raper et al. (2000) show that the turnover timescale can be expressed by,

$$\text{Dynamic Adjustment Timescale} \propto \frac{\text{total volume}}{\text{accumulation rate}} . \tag{1.1}$$

However, when the climate changes, the mass balance changes immediately, so the dynamic adjustment timescale explains why ice-masses lag, to varying extents, behind climate change. This is important to modelling of sea level because people are interested in what happens over the next hundred years rather than the equilibrium change which would take Greenland thousands of years and Antarctica even longer. For instance the accumulation over Greenland is roughly equivalent to that over all the glaciers and ice caps in the world. However, Figure 1.4 shows Greenland's volume to be far greater, and thus the dynamic adjustment timescale for glaciers and ice caps is much smaller than that of Greenland. This means the glaciers and ice caps will reach find a new balance level quicker than the ice sheets.

If climate perturbations over the glaciers and ice caps of the world were comparable to those over ice sheets, then the changes in accumulation over each would also be comparable. Since accumulation over glaciers is roughly equal to that over Greenland, and accumulation is one half of the mass balance, the deficits in mass balance would be comparable across the two. This means that in the short term, before dynamical changes occur, the two forms of ice-mass would give similar contributions to sea level rise, despite the much greater size of the ice sheets. In addition to this both ice-sheets are in very cold climates. This means that the amount their mass balance will react to changes in temperature is very low[3]. In the case of Antarctica this is as good as zero and for Greenland also small. This means the ice sheets actually contribute less than the glaciers and, in fact, Antarctica contributes negatively.

The nature of climate change will accentuate these changes. Warming is projected to be non-uniform, with the greatest increases in temperature taking place in the high-latitudes, especially in the northern hemisphere, where most glaciers reside. This is thought to be because the reduction in snow and sea ice cover at high-latitudes causes an ice-albedo feedback (section 2.1) and because the southern hemisphere has a much larger proportion of ocean which will absorb some of the warming. The precipitation field is similarly variable, with most projections showing the largest increases occurring in the tropics, a decrease in the sub-tropics and then

[3] explained in more detail in section 2

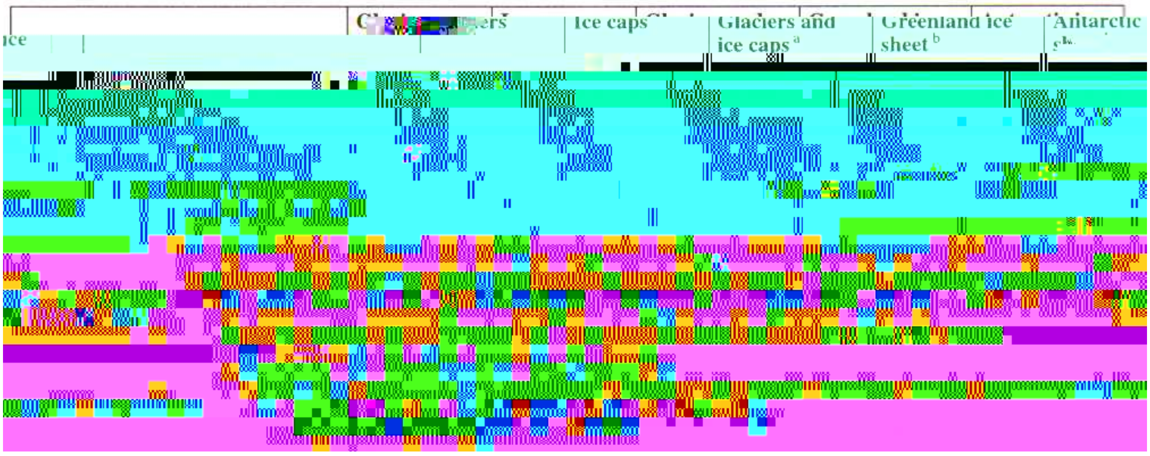

Figure 1.4: from the IPCC's third assessment report, chapter 11

a small increase at mid and high latitudes (Noda and Tokioka (1989), Murphy and Mitchell (1995), and Royer et al. (1998)). The decrease in the sub-tropics is attributed to increased tropospheric stability in the warmer climate.

The increased precipitation levels, expected to average 2 per degree of any temperature increase (Van der Wal and Wil), will not be su cient to maintain balance. Mass balance modelling suggests that for a glacier to stay in balance it would need a 20 to 35 [4] increase in precipitation per 1 °K rise. On the global scale, temperature increases are likely to dominate the balance by far, which means that the total glacial melt rate will continue to rise.

Current projections for the contributions of glaciers and ice-caps to sea level rise, shown in Figure 1.3 suggest a figure of 0.16m over the next century, but with an error margin of 40 . This margin, coupled with the importance of sea level rise estimates to populations worldwide, motivates an investigation of possible improvements to this estimate.

Van der Wal and Wil (2001) show that using mass balance models without taking account of changes in surface area leads to an overestimate of sea level rise by 19 . Whilst their method is an improvement on others, it is still limited in two kew which shoc imp impro

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Dynamical models offer a solution to this situation, but have been previously excluded because the scale at which they run is too large, and when brought down to a scale suitable for glaciers they then have the same problems with lacking sufficient meteorological data. However, the development by Marshall and Clarke (1999) of a subgrid parameterisation that captures the detail of the terrain and its consequential meteorological effects allows dynamical models to be used for global glacier volume estimates for the first time.

Chapter 2

Ice-mass Modelling Theory

Accu[illegible]tion [illegible]ersus A[illegible]tion

Precipitation onto an ice-mass may either land as snow/hail, or rain. The part that falls as snow/hail, and the part of the rain that froze upon impact may survive a whole year to become firn. It may remain in place and go through several further stages of compaction under subsequent layers of precipitation[1], until it reaches the necessary pressure ($830 kg\, m^{-2}$) for its internal crystals to align to become ice. Alternatively it may melt on the surface. This melt water, along with the rain that did not freeze, may then evaporate or run-off; or it may soak into the snowfall and refreeze. The accumulation in a given year comprises the part of the precipitation that did not evaporate or run-off.

[1] [illegible]

point of ice and β is derived from the clasius-claperyon equation. Basal flow is important in ice equation.ice sheets as to most other surfaces on Earth. This is important because in forest terrain, sunlight, which is shortwave radiation, will be absorbed by the ground and heat, which is longwave radiation, will be emitted; this warms the atmosphere. Over an ice mass, however, most shortwave radiation is reflected, due to its high albedo, and so there is less heating of the atmosphere. This is a positive feedback process which means it encourages whatever the ice mass is doing. In periods of growth the atmosphere is cooled which promotes further growth, whereas in periods of retreat the reduction in the albedo-effect warms the atmosphere, encouraging further retreat. Ice albedo decreases down a glacier and through the course of the summer melt season.

The problem with using energy balance models on a global scale is that the detailed information of surface fluxes is only available for a few glaciers worldwide. Furthermore, the climate models that are used to generate projections of future climate scenarios do not produce this level of detail. For this reason, temperature index methods are used. These parameterise ablation and the fraction of precipitation to fall as snow as functions of temperature[2]. This reduces the problem of knowledge of climatic conditions to knowledge of temperature and precipitation, which can currently be modelled on a global scale at a resolution of $10 - 100\,km$.

2.1.1 Mass Balance Sensitivity

An ice sheet starts life as a glacier; effectively they are the same thing, with the same rules, just applied on a different scale. It has been established that energy balance modelling generally requires more information than is currently available, and that temperature index methods allow us to limit our dependence to temperature and precipitation modelling. On a global scale the approach of nesting regional climate models in general circulation models (Hostetler and Clark, 1997) and downscaling of GCM output (Glover, 1998) has improved the realism on the $10 - 100km$ scale. However, glacier models at present require a level of detail that could not function with this little detail. Moreover, of the +160,000 glaciers in the world, only 100 have been mass balance records greater than 5 years, and only 40 have records longer than 20 years.

One way around this is to group glaciers by their mass balance sensitivity - that is how much their mass balance changes per degree change in temperature, and so maximise the knowledge we have. Mass balance sensitivity varies across the globe; summer temperature increases are generally the most important increases for glaciers because the winter temperatures are suf-

[2] More details on temperature index methods are given in section 3.2

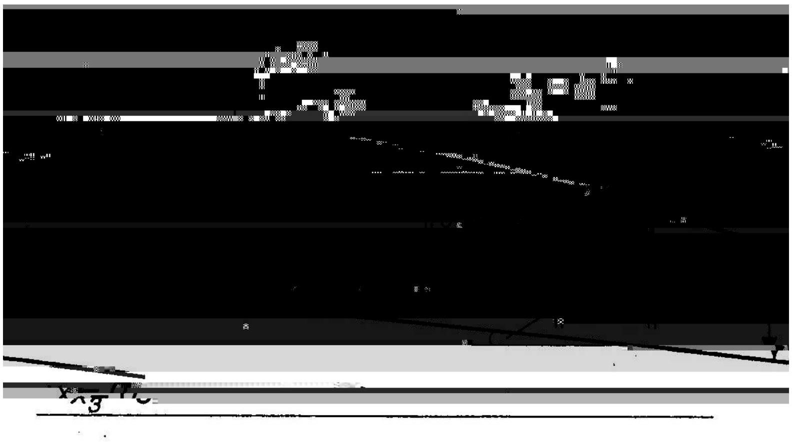

Figure 2.1: from The UBC Ice Sheet manual

ciently low that increases do not cause melting. However, for low-latitude glaciers the annual average rise is more crucial. The problem this runs into is that whilst we have a good idea of the area and volumes of glaciers in these separate regions, we lack accurate measures of the specific mass balances.

ce Thic ness the rst prognostic N N e

The spatial and temporal evolution in thickness of an ice mass gives an easy to grasp measure of what the ice mass is doing and as such is one of two key prognostic outputs used in most contemporary models, the other being ice velocity.

The diagram above shows a simple ice sheet profile, with h^{l}_{a} representing the ice sheet surface, h^{B} representing the bed, and 0 taken as a large scale base marker, say a geoid of the earth, or sea level. The thickness of the ice sheet is given by $H = h^{l}$ ¡ h^{B}, and the mass balance equation gives the evolution of the sheet iceD7-1 0 Td (The)64Td 07 0 Td (profile,)Tj 3t.85068 0 Td 18.bne

ce DynN ics

Ice can flow by internal deformation or by basal interactions. The rate of flow is variable, but can be segmented into two main forms, stream flow and sheet flow, with stream flow an order of magnitude or more greater that sheet flow.

2.3.1 Sheet Flow

If we consider an example where our ice-mass is frozen to its bed, as is the case for glaciers most of the time, then the velocity is zero at the bed, and sheet flow is the internal deformation of the ice; the plastic response of the ice to the pull of gravity. The deformation, known as viscous creep deformation, occurs as ice crystals slide over one another and change shape (see PTheanotheri99glaciers8another

CHAPTER

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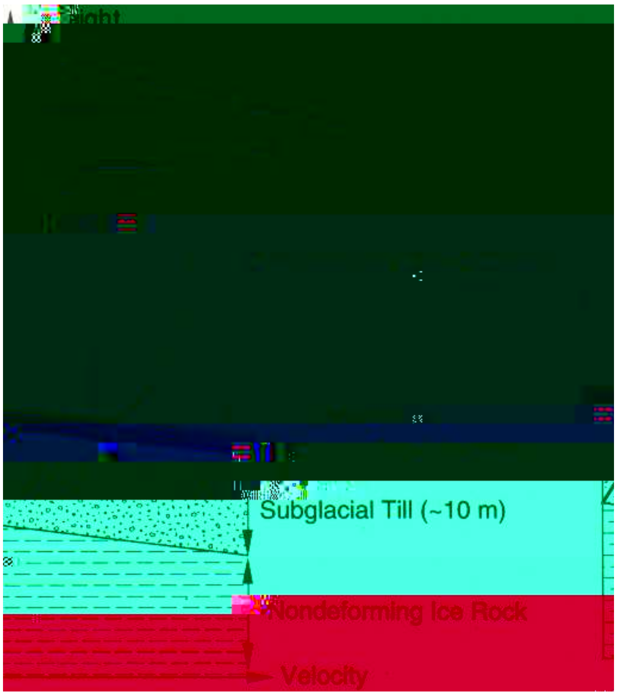


and instead will form a layer between ice and rock. This decouples the ice mass from its bed, and reduces the basal drag massively. This allows for greater ice flux, which may cause greater basal friction, and the temperature rise in turn generates more water from melting. Where the bedrock is not solid, but not all the water is absorbed, the lowered friction generated by the water causes some increase in ice motion. It also causes some deforming of the bed and it this that causes glacial till and moraines. The physical processes involved are examined in greater depth in Patterson (1994, chapters 7 & 8), and are illustrated in Figure 2.3.2.

When modelling ice stream flow the shallow ice equations become insu cient, as the longitudinal and transverse stresses become more influential than the vertical stresses. Alternatives to the shallow ice approximation have been used by Pattyn (2003), Saito et al. (2003) and Marshall (1999). Pattyn and Saito retain the longitudinal and transverse stresses in a so-called "higher-order" model, solving using the force-balance 23.2404 0 Tdfqc introduces a parameter $\alpha(\lambda, \theta, t)$, which is the areal fraction of ice streams in a control area. The equations for thickness, ice temperature

The temporal evolution of the rate of ablation, and the consequential ability to model the stabilisation of a currently retreating glacier, are only possible with the use of a dynamical model. As part of their mass balance modelling Van der Wal and Wil (2001) used a parameterisation of,

v

Payne speculates that there may be another factor also causing the effects seen in Figure 2.2 - the coupling between thickness and flow. As ice velocities increase, the same flux of ice can fit through a smaller thickness, thus surface elevation falls. However, ice on the surface follows the surface, rather the topographic, gradients. Thus it will move towards the lower surface of the fast flow, and help to increase the velocity further.

At present it is still not certain that the results of the EISMINT study do reflect real processes and are not due to numerics or the simplified geometry utilised.

Chapter 3

Model Parameterisation Schemes

Chapter 2 showed that inclusion of dynamics is important to improving the skill in modelling ice masses. However, dynamical models have not been used in global glacier studies before. This is because the climatological and topographic data required for a global model lacks the necessary spatial resolution for individual glaciers.

This chapter examines the consequences of a lack of spatial resolution and illustrates a method for improving this as well as explaining the degree-day scheme 3.2 that is used to calculate the mass balance.

Nuc eNtion ithin ce Sheet Mode s

Inception means the development of any ice, from a blank starting field, whereas nucleation refers to the early development of an ice mass; this could occur from nothing, but is more likely to involve the evolution from small, existing glaciers.

Consider a glacier in balance, located in a region of high relief, and apply a climatic perturbation such as a period of increased precipitation or lower temperatures. This encourages the

pres3780ie Td (e)Tj 13.7771 0 Td (lo)Tj 8.17036 0 Td (w)Tj

precipitation, a decrease in temperature from -5.342° by just hundredths of a degree is sufficient to move from the first signs of ice to a $2000\,m$ thick ice sheet. Similar non-linear sensitivity to the climatology is visible in the precipitation; for -8° an increase from $0.48\,m\,a^{-1}$ to $0.49\,m\,a^{-1}$ is all is needed. The consequences of this are that a small error in the climate data will have a very large effect on ice development. In reality mass balance varies spatially, and climatic conditions vary temporally, so that the onset of nucleation would not occur so abruptly.

Ice sheet models have been run as fine as $20-25\,km$ (Huybrechts, 1996) but are limited by the resolution of the climatological data that is used. In global models the grid cells are $\approx 100km$, at which scale individual valleys, and peaks are not resolved, and whole mountain ranges may be represented by only a few grid points (Alps, Pyrenees). The ice sheet models either miss the accumulation at high-elevations or fail to account for low-elevation ablation, which affects the generation of ice sheets in the correct places (Rind et al 1989, Marshall and Oglesby 1994).

Marshall and Clarke (1999) suggest that it is this lack of topographic detail that causes the lack of realism in studies of the growth and decay of past ice sheets. They developed a parameterisation that captures some of the subgrid topographic detail to improve the quality of inception modelling. However, by improving the treatment of accumulation and ablation they make it possible for the first time for an ice sheet model to be used on a global scale for glacier modelling.

The details are left to and

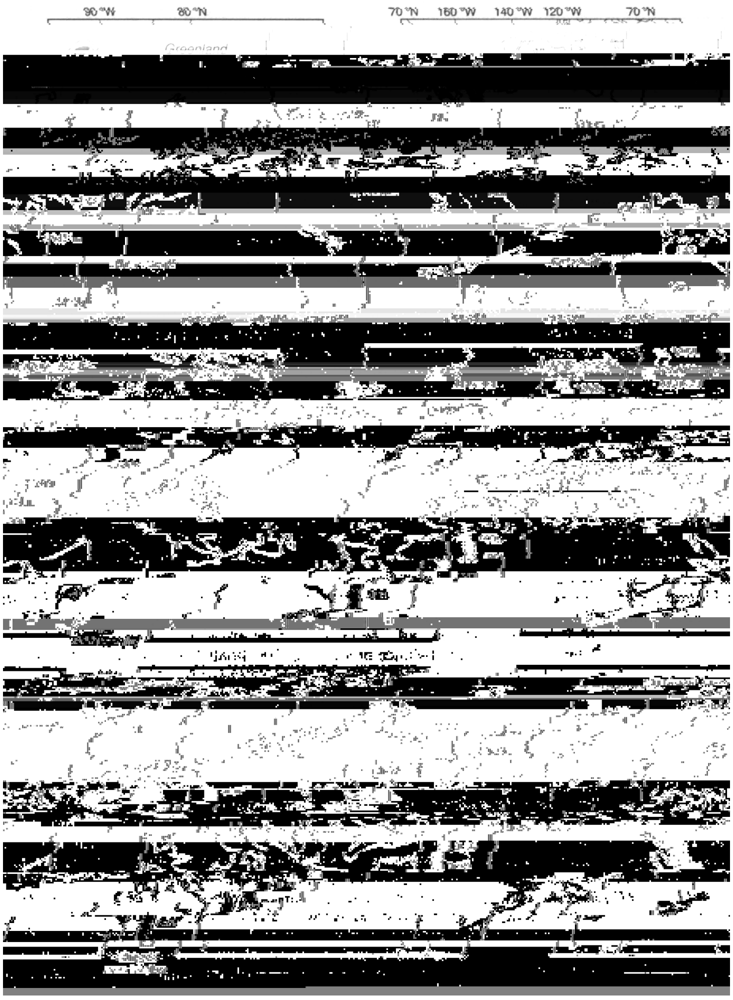


Figure 3.2: **Marshall and Clarke (1999); Top row is the Canadian Arctic, bottom is the North American Cordillera. The 1st column shows present day ice, the 2nd shows the large scale model's ice thickness output, and the 3rd shows areal coverage with the subgrid parameterisation.**

of ice sheets, but this is not possible as the level of accuracy required would be too expensive to

The melt-rate of snow is given by,

$$\dot{m}_{\textsf{snow}}(\lambda, \theta, t) = d_{\textsf{snow}} PDD(\lambda, \theta, t)(1 - r_{\textsf{f}}), \tag{3.4}$$

where $r_{\textsf{f}}$, the refreezing factor, accounts for water that melted but then refroze rather than running-o .

Ice ablation is then the result of any surplus heat energy after the snow has been melted. It is calculated from the numow

Longitude	$71.25\,^{\circ}E$	$75\,^{\circ}E$	$78.75\,^{\circ}E$	$82.5\,^{\circ}E$	$86.25\,^{\circ}E$	$90\,^{\circ}E$	$93.75\,^{\circ}E$
Maximum(m)	6330	8311	6970	7010	6184	6480	6055
Mean(m)	2388	3657	5158	5133	5063	5026	4629
Minimum(m)	253	513	2466	2406	4705	4012	3032

Table 3.1: Di erences between

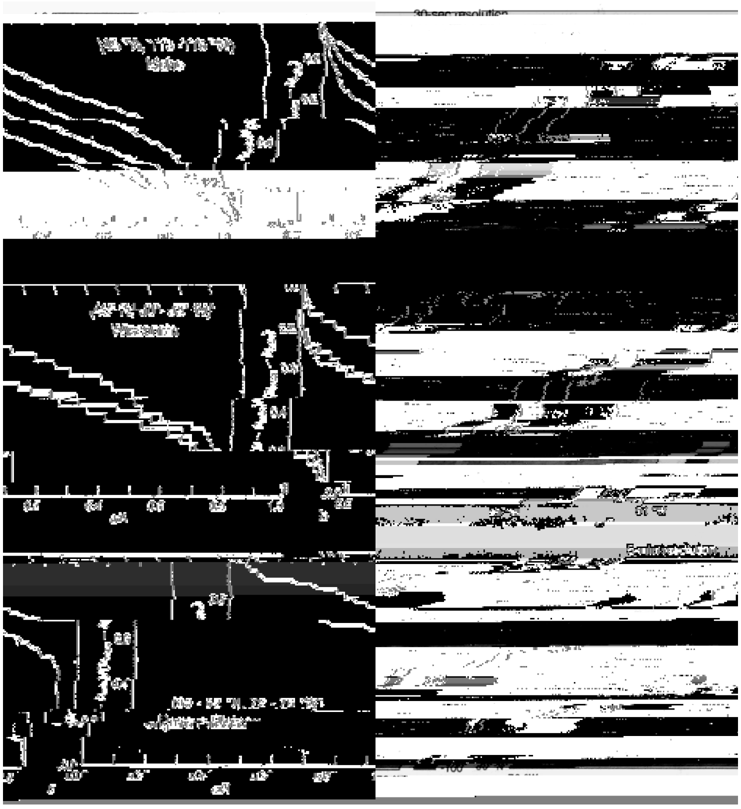


Figure 3.3: from Marshall and Clarke (1999)

mapping,

$$\frac{h^{B} - h^{B}_{\min}}{h^{B}_{\max} - h^{B}_{\min}} = f(a/A), \qquad (3.6)$$

where A is the grid cell area and a/A represents the cumulative subgrid area above elevation h^{B}. The synthetic hypsometry curve used is then given by,

$$\hat{h} = \frac{h^{B} - h^{B}_{\min}}{h^{B}_{\max} - h^{B}_{\min}}$$

$$= \Big[\mathrm{atanh}[b(2a/A - 1)] + \mathrm{atanh}(- \quad R \quad Tf \quad Td$$

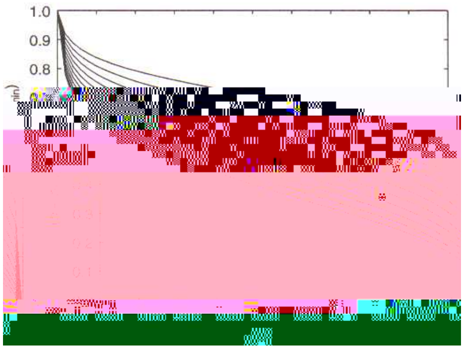


Figure 3.4:

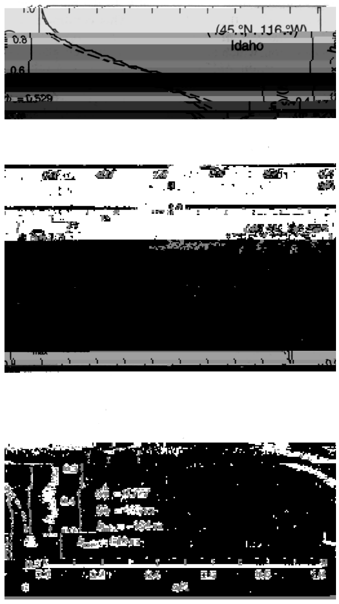


Figure 3.5: from Marshall and Clarke (1999)

elevations of the bins evenly over the range, with spacing Δh

$$h_k^B = h_{min}^B + \Delta h \left(h - \frac{1}{2}\right), k \in (1, n_h). \tag{3.9}$$

• A rearrangement of equation (3.7) then gives the subgrid cumulative area,

$$a/A = 1/2 + \frac{\tanh(\hat{h}^{-c} - 1/2)}{2b}, \tag{3.10}$$

which is easily broken down into the subgrid area for each bin.

At this point, for each bin, subgrid elevations and areas have been found. Accumulation ($\dot{a}_k$) and ablation ($\dot{m}_k$) are calculated using the same degree-day parameterisation that was explained in section 3.2, but with improved temperature and precipitation estimates.

Temperatures for each bin are generated by adjusting for elevation above sea-level using the adiabatic lapse rate. This is a common feature in dynamical models because it allows them to evolve the surface temperature with changes in ice sheet thickness. However, rather than adjusting for the mean cell height the temperatures in each bin are computed based on the elevation of that bin. Assuming that the precipitation data is at surface height rather than sea level and so already shows the effects of elevation, it is spread evenly over all bins except above a threshold known as the drying height.

This threshold is a chosen distance above the mean cell level, designed to mimic the elevation-desert effect at high altitude or when an area becomes heavily glaciated, where the air becomes very dry and thus precipitation drops off. Above this point the precipitation is reduced exponentially with height. In cases where the drying height is reached, and upper levels have their precipitation reduced, the surplus is then spread evenly over the lower levels. When modelling for North America, Marshall and Clarke (1999) chose to use 1000m as their drying height.

Marshall and Clarke (1999) recognized in their original work that the processes involved in precipitation distribution are very complex, especially so in mountainous regions, and that the details of absolute elevation, local winds and moisture sources may be more important than the simple elevation effects embodied in the drying parameter. Barry (2001) presents work by Lauscher (1976), that is repeated here in Figure 3.6, when discussing the details of precipitation lapse rates. They are noticeably different for regions of interest in this study; middle latitudes have significantly greater precipitation with height whilst polar regions have decreasing precipitation linearly with height. Barry notes that whilst Lauscher's profiles are useful generalities, the local and regional complications often outweigh these simple rules, with some mountains having different profiles on different slopes (Mt. Cameroon, Lauer (1975)) and others having seasonal fluctuations (Bavarian Alps, Erk (1887)). This indicates that Marshall and Clarkes (1999) precipitation lapse rate[4] may be incorrect when applied to a global mapping.

Accumulation and ablation have been accounted for, now dynamics need to be applied. Recall that the ice build-up on the hypsometric curve is an idealized distribution of the areal coverage of ice across the elevations within a cell, and not a picture of the true, often complex, subgrid topography and ice distribution. Thus the dynamics as described in section 2.3

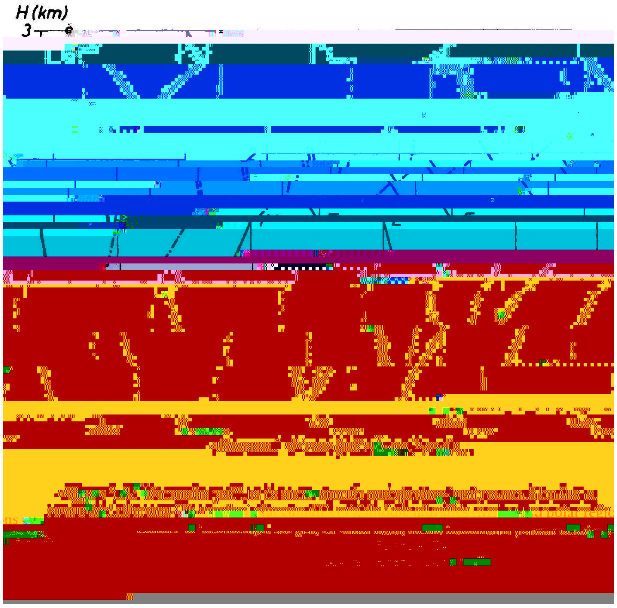


Figure 3.6: from Barry (2001), p233

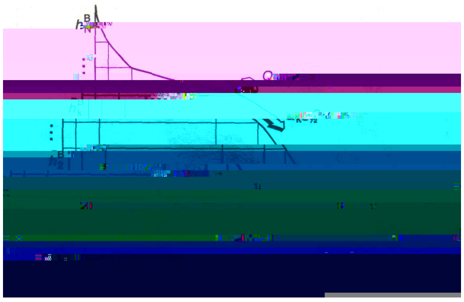


Figure 3.7: from Marshall and Clarke (1999)

flux of ice is then $a_k Q_k^{m+1}$.

Ice is assumed to be incompressible. Then for all bins except for the top (where there is no influx of ice from above) the ice thickness evolution in bin k is the result of the incoming ice flux from bin $k+1$, the outgoing ice flux from bin k, and the mass balance of bin k,

$$\frac{\partial H_k}{\partial t} = \frac{a_{k+1}}{a_k} Q_{k+1}$$

Similar to a half-life, $L_{\mathbf{k}}$ is a length scale that measures how long ice stays at one level. A first order Taylor series expansion of equation (3.12) gives $Q_{\mathbf{k}} \propto H_{\mathbf{k}}(\Delta h_{\mathbf{k}}^{l\,m}/L_{\mathbf{k}})^3$. Marshall explains that they chose an exponent of 3 in equation (3.12) because the dependence on surface slope is then similar to Glen's flow law for glacier ice (equation (2.6)),(Paterson, 1994). He goes onto suggest an appropriate form for $L_{\mathbf{k}}$ in terms of cell hypsometry,

$$L_{\mathbf{k}} = \frac{L_0 a_{\mathbf{k}}}{A}, \tag{3.13}$$

where L_0 is a horizontal length scale for the cell. The perimeter of the cell should not be used for L_0 because cell size will vary depending on (1) the chosen grid resolution and (2) on meridional position if on a spherical grid. If it were, then the flow of ice would be independent on the discretisation chosen. Marshall and

over. The data we will be using works from the Greenwich meridian eastwards. This will mean the loss of the small glaciers in France and Spain, as well as some of the Norwegian network. However, this loss is minimal, and the ice sheets at the poles are not of direct concern to this study, thus Marshall and Clarke's approach is maintained. As a sub-note, the model encounters convergence issues associated with the very small grid spacing when extending to the poles ($90\,^{\circ}N/S$), and so we choose to run from $85\,^{\circ}N$ to $85\,^{\circ}S$.

Chapter 4

Design of Experiments

There are two components to consider in designing the model runs: examining the data to be used, and justifying the experiments to be run on those data.

The nput Data

The UBC ice sheet model requires, as a minimum, climate forcing, bed topography and initial ice thickness. However since this study will be using inception runs (section 3.1) the initial ice thickness field is

Topographic data that is required is the maximum, mean, and minimum elevations for each grid-cell, and this is computed from the etopo2 DEM data, which was provided at 2-minute resolution. The land fraction has also been calculated with the etopo2 data, and gives a subgrid representation of the land fraction of each cell. The mean elevation and land fraction are used by the model as the grid-scale bed topography. The mean elevation is also used with the maximum and minimum elevation data to generate the hypsometric curves for each grid-cell. These curves are then used to calculate subgrid bin areas and elevations.

4.1.2 Climate Forcing

The atmosphere and oceans are modelled separately, with the models coupled to each other through sea–surface fluxes of heat and carbon. Subgrid processes such as cloudiness and precipitation are parameterised in the atmospheric model. They are limited in their ability to model the climate [2] but they are, however, the only physically based way to calculate a regional and seasonal pattern of climate change. This is important because climate change will not be uniform and it has already been shown in this study that glaciers react differently to summer and non-summer changes.

The climatological data consists of temperature and precipitation grids. Two versions will be applied: a modelled and an observed dataset. The modelled dataset is from the 3rd version of the UK's Hadley Centre Climate Model (Gordon et al. (2000)) and is provided at a resolution of 2.5 ° in the latitude, 3.75 ° in the longitude. The climate model (hereafter HadCM3) is run with an atmospheric composition approximating that of 1860 for a period greater than 1kyrs, during which it is stable. The data is the result of a 100 year mean from within this period.

Most ice sheet models are designed to run on either annual or monthly climate data, the UBC model will accept either, but as was established in section 3.2 monthly data are used due to the greater accuracy. Data that can distinguish between the seasons is especially important if a model is to be used in climate change experiments, as change is most likely to be both regionally and seasonally varying (section 3.2).

HadCM3 produces surface-height temperature data, whereas the model requires sea-level temperature data. This is because it will adjust the surface temperature with surface height, which can vary due to changes in ice thickness[3]. This means that the HadCM3 data must

[2] see Gordon et al. (2000) for details of the problems with the Met Office's Hadley Centre model

[3] on much longer timescales beyond the scope of this study the ground thickness may also change due to

be adjusted, and since the topography is dictated by etopo2 DEM data, the temperature data must be adjusted

Experiments Undertaken

The experiments fall into three broad categories; implications of the hypsometric parameterisation, equilibrium runs, and climate change scenarios.

4.2.1 The Hypsometric Parameterisation

The model allows for the hypsometric parameterisation not to be applied. This means the effects of the parameterisation can be examined, by quantitative comparisons of the total ice volume and total areal coverage, and qualitative analysis of the areal coverage. Qualitative analysis of areal coverage involves examining plots of the areal coverage to see if the separate model runs produce ice in the same areas, and more importantly if the parameterisation improves the skill of the model. This can be measured by comparing the output to the current spread of ice across the globe. To do so would be to reproduce, and thus reinforce, a result Marshall and Clarke (1999) found when modelling the North American continent.

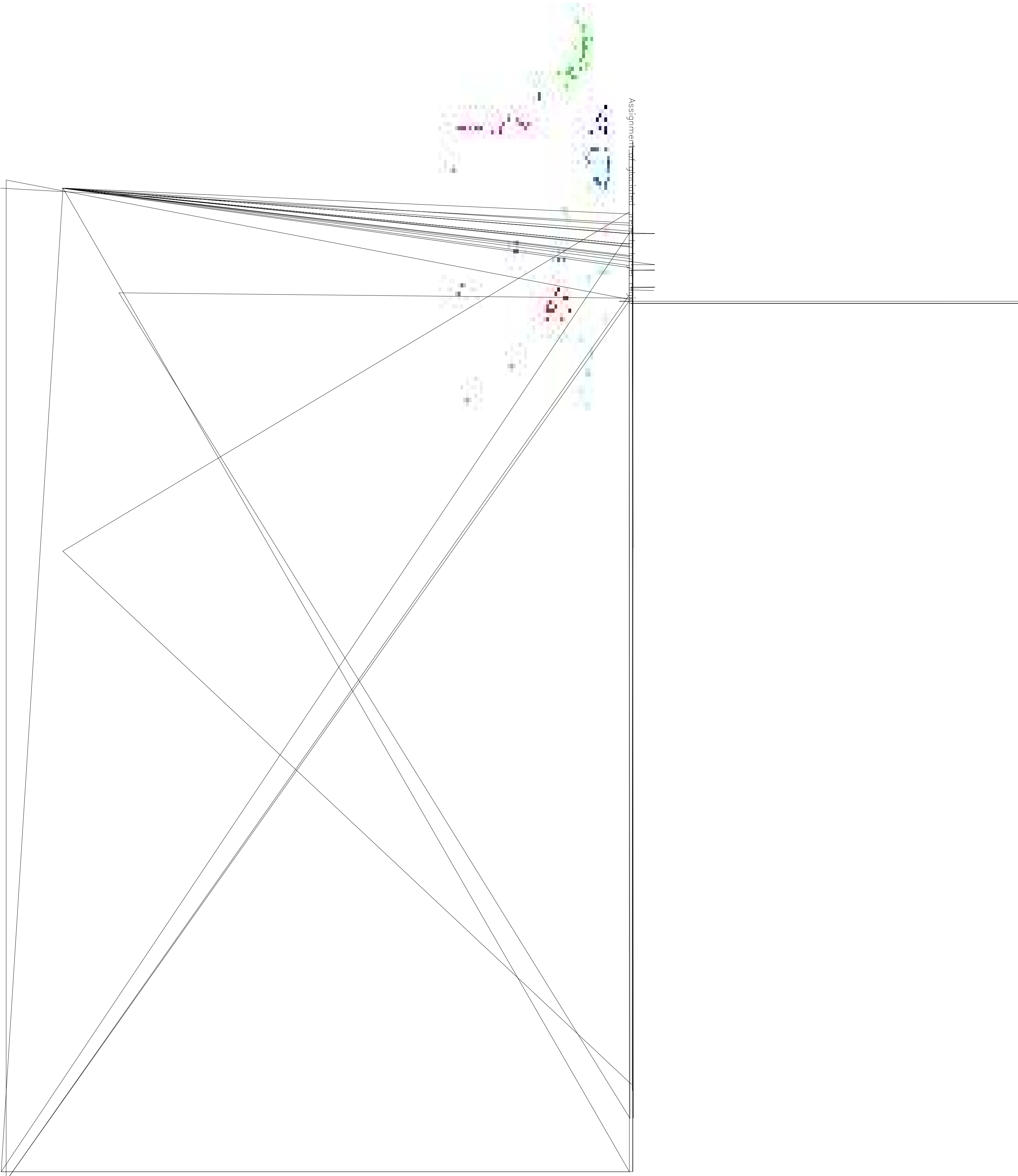
Examining the subgrid ice thickness in a variety of bins at a limited number of specific locations shows if the model is distributing the ice across the heights appropriately. Moreover, choosing suitable locations that have distinct hypsometries whilst similar mean altitudes will test another of Marshall and Clarke's results; that distinct hypsometries produce distinct ice-masses.

4.2.2 Control Runs

These runs use the subgrid hypsometry, and test to see if the model will produce a stable and realistic state for glaciers. They will be 1500 year runs, which are too short to properly develop ice sheets, but should give glaciers and ice caps sufficient time to develop. Most glaciers have dynamic adjustment timescales in the range 10-100 years, whilst ice caps are a little slower, nearer the 800-1000 year mark. We test for convergence to an equilibrium level by examining a time series of total areal coverage and total volume.

Should an equilibrium be reached, use of the two different datasets, in combination with regionalised glacier volume data[5] (see Figure 4.1) will allow for a comparison of the modelled volumes with observationally based inventories.

[5] Zuo and Oerlemans dataset with an der al and ils data for Greenland



winter season.

Such simple climate changes are not likely to represent the genuine climate change, but they are significantly easier to comprehend. They aid the understanding of how glaciers on a global scale will react to climate change, and their sensitivity to some of

Chapter 5

Results

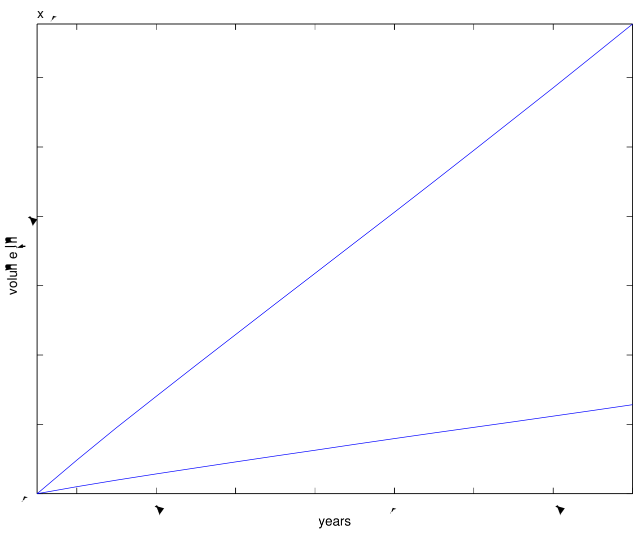
As explained in chapter 5 there are three classes of experiments considered in this study. The results of the control experiments for the HadCM3 and CRU datasets are considered first as these will be used as a benchmark in the other tests. Throughout this chapter figures are constructed such that the HadCM3 run is shown in (a) and the CRU results in (b), unless otherwise noted.

Contro Runs

The first consideration with the model is whether it develops a stable state. The totals of volume for the two datasets are shown in Figures 5.1(a) and (b). The total global ice volume, inclusive of the ice sheets is shown in 5.1(a).

The ice sheets take much longer to equilibrate with their climate because of their longer timescale for dynamic adjustment (see section 1.4), and have much greater volumes. This explains why the curves in Figure 5.1(a) show no sign of stabilising. The lack of Antarctica in the CRU dataset partly explains why there is less total ice volume.

Figure 5.1(b) shows that in both runs glacier volume converges. Noting the di erence in scale, the HadCM3


x
years

	Estimate	CRU	HadCM3	Improved
Ice Area ($10^6\,km^2$)	0.598	0.4	2.3	0.99
Ice Volume ($10^6\,km^3$)	0.218	0.165	0.315	0.24
SLR equivalent (m)	0.603	0.052	0.998	0.68

Table 5.1: Comparison of the volume and area totals produced by the model, under the two datasets, to the totals from an observationally based dataset.

the HadCM3 model gives better results for volume, but over-estimates the area. This tendency can also be seen across most regions, AppComparison(area)4T0823810.9091 Tfb)illustr72 4.04j 134639.8165 0 Te two pairs, marked in colours by pairing, in Figure 5.5. The cell pairings that were chosen are such that the di erence in their mean elevations is less than $30m$. Unfortunately the Himalayas are not resolved in the run without the subgrid hypsometric parameterisation and so a comparison of the cells with that model can not be made, except that the temperature at the mean altitude is su cient to melt all accumulation when considered on a grid scale.

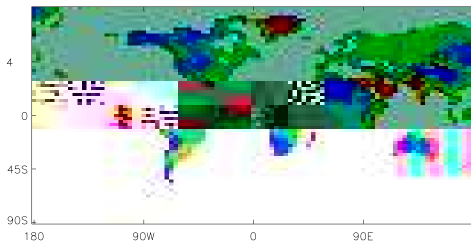
Figure 5.5 shows the thickness of ice that has collected in each subgrid bin. Only one of the cells collects any ice at the mean elevation, most likely due to a strong downward flux caused by a larger range in this cell. This figure is illustrative of several key features of the subgrid hypsometric parameterisation.

Firstly it shows that ice develops in a reasonable way on a subgrid scale: it is not a single anomalous bin that is collecting ice but a series of bins. Secondly, the distribution of subgrid thickness is similar to reality, where thickness increases downslope until an area of large melting occurs. Note that the di usive relaxation used in parameterisation will exaggerate this slightly. Thirdly, cells which have very similar mean heights and climatic conditions are developing quite di erent ice thicknesses and distributions because of their subgrid terrain. This directly supports work by Marshall and Clarke (2001), showing thegrid tst,l.

cell, the control run stabilises at around $0.6m$, with the rest of the other runs spread as they are now, but simply closer together.

The aim of a sensitivity study is to examine how the model responds to changes in its dependency fields, in this case precipitation and temperature. Whilst the figure may be polluted by the Alaskan cell the trend is still clear; lowering the climate by one degree has a similar effect to raising it so there is symmetry. The changes, however, are not linear an increase of $2\,^{\circ}C$ and 4 in the precipitation field (twice the standard change) does not decrease the volume of glaciers as much as the $1x$ change. The model does not seem to be far from a linear decrease in the $2x$ case so more work is required to clarify this.

As temperature increases, the global glacier sensitivity to temperature appears to decrease. It is not clear whether the same can be said for precipitation, if not then precipitation effects may well become more relevant in the reference, Church et al. (2001) estimated the contribution to sea level rise from glaciers and ice caps would be $0.5\,mm\,a^{-1}$.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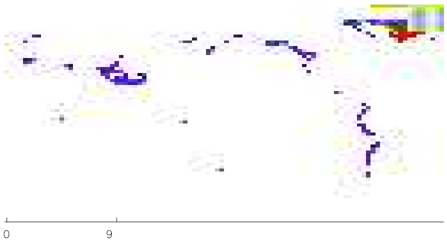
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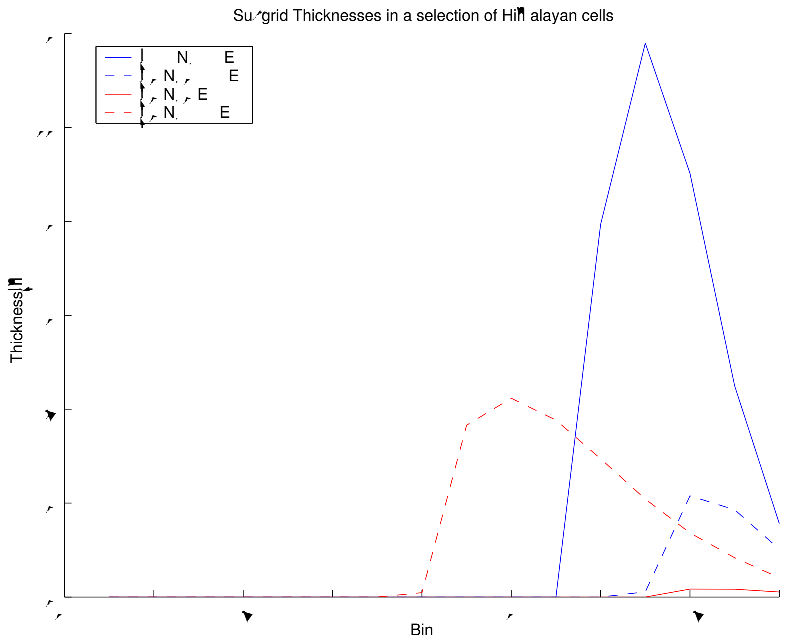
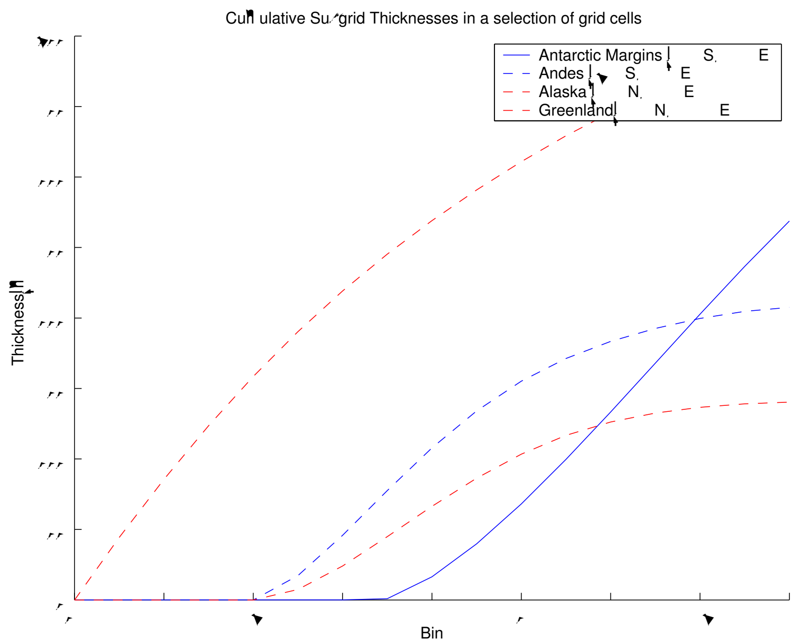
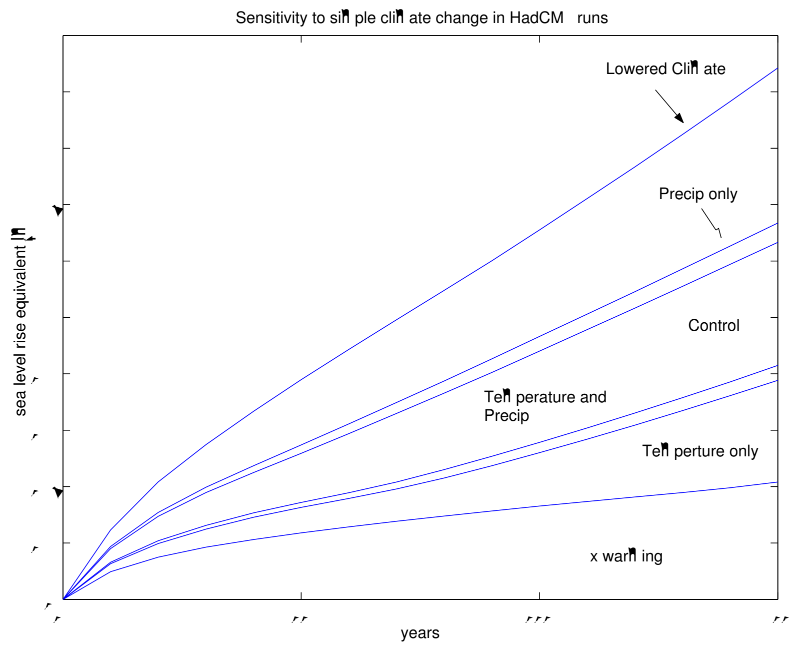
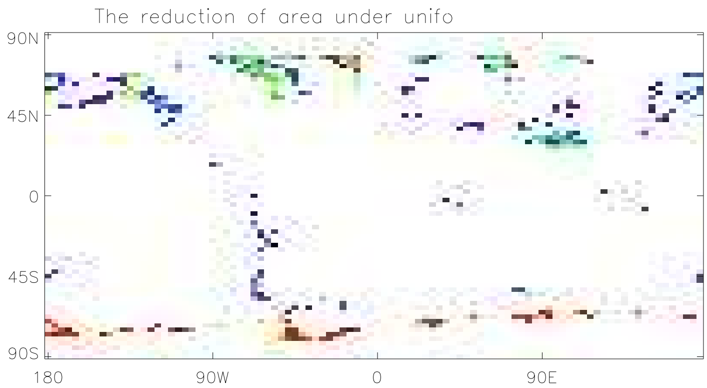


Figure 5.5: From the HadCM3 control run


ulative Su grid Thicknesses in a selection of grid cells
Antarctic Margins
Andes
Alaska
Greenland
Thickness
Bin


Sensitivity to simple climate change in HadCM runs
Lowered Climate
Precip only
Control
Temperature and Precip
Temperture only
x warming
sea level rise equivalent
years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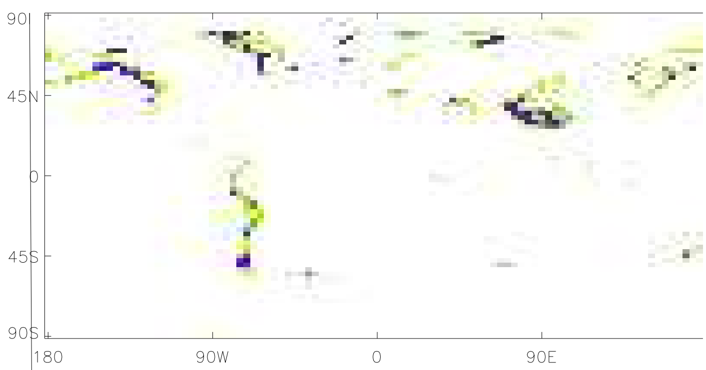
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Chapter 6

Conclusions

Summary

The key findings of this study can be summarised in seven points.

1. Inclusion of subgrid topographic detail is essential in order for an ice sheet model to be able to reproduce glaciers on a global scale. The ability to model glaciers accurately with ice sheet models should improve long-term predictions of their contribution to sea level rise.

2. Marshall and Clarkes (2001) subgrid hypsometric parameterisation enables an ice sheet model to reproduce realistic equilibrium states for glaciers. There is much interest in the transient change over coming centuries, as it will be the change that we all feel, but planning for the long-term future requires policy-makers to have an idea of what climate we'll be living in in the future. Whilst this model was not originally designed for modelling equilibrium states, it has been shown to be useful in their consideration.

3. The model is sensitive to the climatology applied. Until HadCM3 or CRU is shown to be the "better" model, there is low confidence in the magnitude of sea level rise predicted in these tests, because of the large range.

4. Under a $4 \times CO_2$ climate change scenario the model predicts that global glacier volume will decrease by (2.1 x 10

5. Glaciers' mass balance is dominated by their sensitivity to temperature changes over precipitation changes. and their relationship

6. If sensitivity to precipitation changes

Appendix A

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Appendix B

Flowcharts

These flowcharts illustrate how particular parts of the core dynamics and thermodynamics are solved.

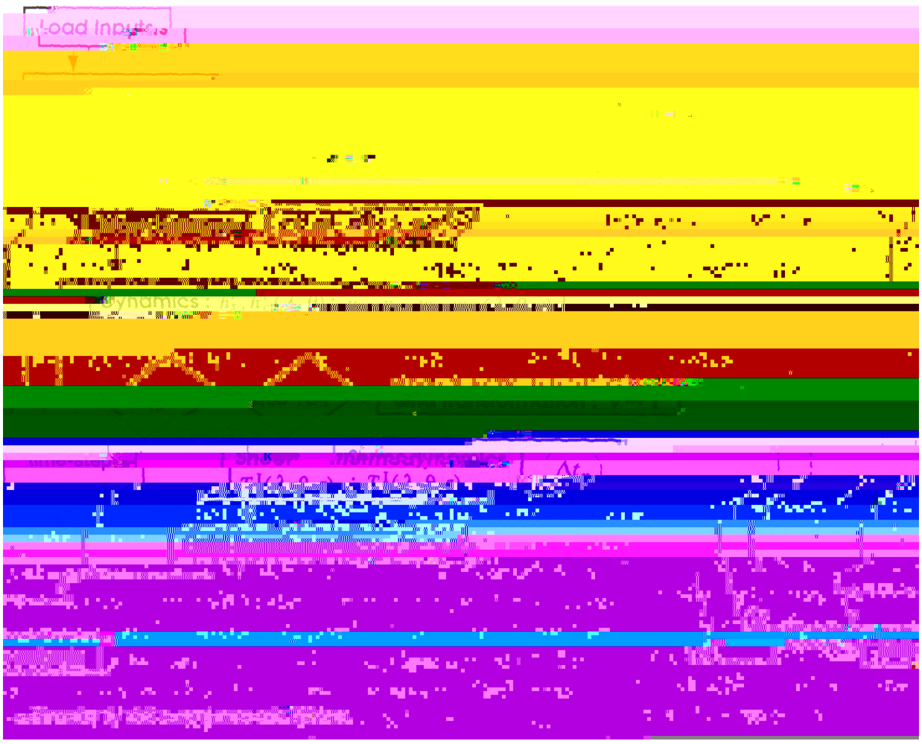


Figure B.2: from Marshall, (1997c).

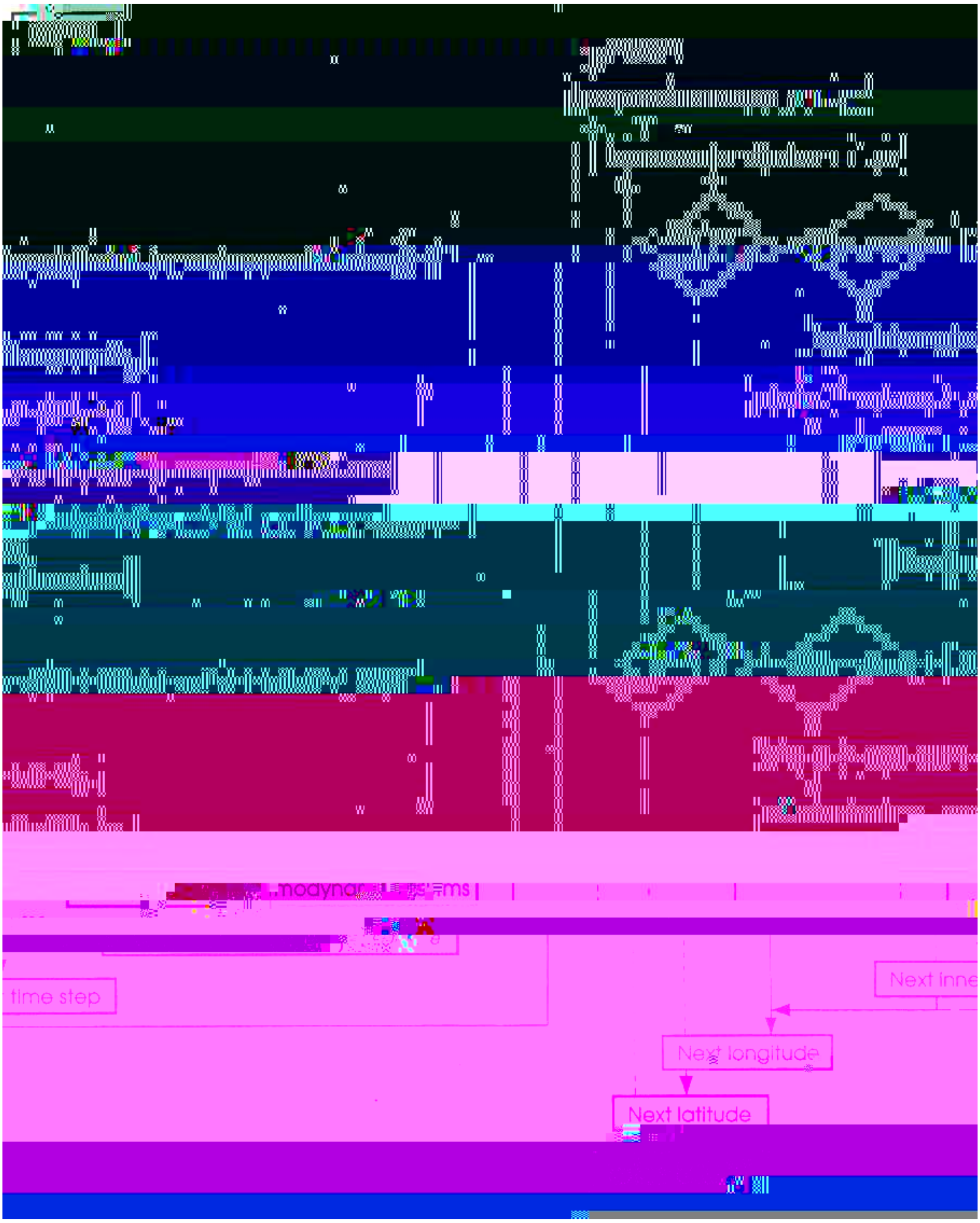


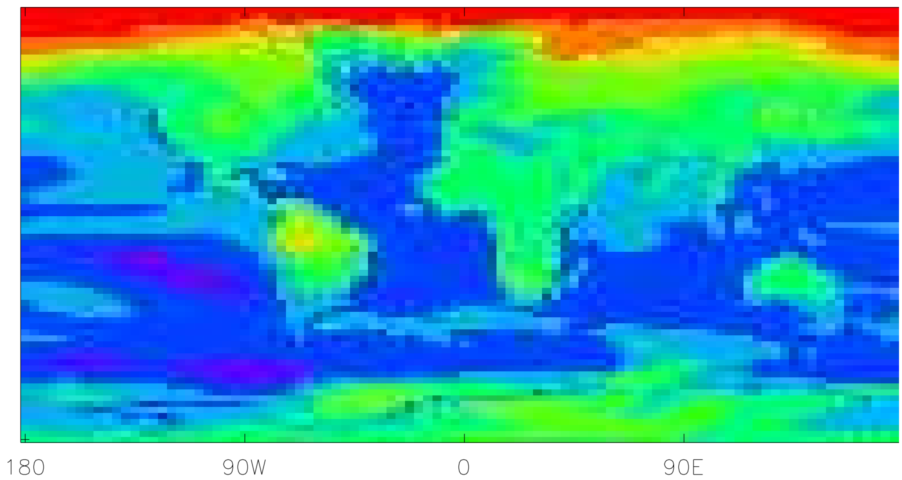
Figure B.3: from Marshall (1997c)

Appendix C

Excel charts

Appendix D

Graphs of HadCM3 Projections of Climate Change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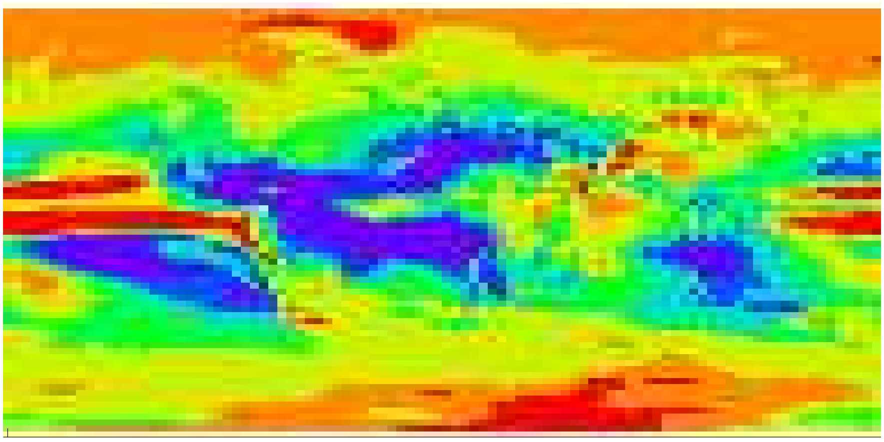

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