Abstract

The attribution of recent global warming to anthropogenic emissions is now well established. However, the relation of recent changes in other properties of the climate system to human activities is not as clearly understood. The aim of this thesis is to improve our understanding of this relation in the case of two of these properties, namely the diurnal temperature range (DTR) and modes of tropospheric variability.

The DTR, the difference between daily maximum and minimum temperatures, has decreased over global land areas at a rate comparable to the mean warming. Model simulations including the effects of human emissions produce a comparable change, albeit of smaller magnitude. This decrease results from increased reflection of solar radiation by clouds moderated by decreasing soil moisture, mostly through its effect on the ground heat capacity.

Recent trends in indices of some modes of atmospheric variability suggest the possibility that forced climate change may manifest itself through a projection onto these pre-existing modes. Model simulations indicate that this is plausible in the case of sea level pressure, but only partly so in the case of surface air temperature. On the interannual time scale examined in this thesis, these projections are consistent with a linear interpretation, rather than a nonlinear one.

These results are, however, sensitive to the representation of small scale processes in the models. For instance, the DTR response depends strongly on the representation of cloud and land surface processes. Further examination of the response of one of the tropospheric modes, namely the Southern Annular Mode which represents the meridional shift of the mid latitude jet in the Southern Hemisphere, indicates that it is sensitive to the parametrisation
of sub-grid scale mixing in the ocean. Nevertheless, these results suggest that the recent changes are consistent with enhanced greenhouse warming, and indicate that they are likely to continue into the foreseeable future.

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Five and a half years ago I was in a bit of a fix.
Some people helped me out.
Thanks.

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You’d probably guessed that anyway. The Census report, like most surveys, had cost an awful lot of money and didn’t tell anybody anything they didn’t already know - except that every single person in the Galaxy had 2.4 legs and owned a hyena. Since this was clearly not true the whole thing had eventually to be scrapped.

— So Long, and Thanks for All the Fish, by Douglas Adams
Chapter 1

Introduction

1.1 Introduction

The prospect of a global climate warming due to anthropogenic emissions, mainly of carbon
dioxide (CO\textsubscript{2}), is one of the most important environmental challenges faced by human
society. The Intergovernmental Panel on Climate Change (IPCC), the United Nations
body charged with summarising and assessing our current knowledge of the issue, projects
a 1.4-5.8°C warming over the next century (Cubasch et al. 2001). This is on top of the
0.6 ± 0.2°C rise since the industrial revolution (Folland et al. 2001). Other properties of the
climate system are also expected to be changing (Cubasch et al. 2001) and indeed recent
climate change has been characterised by more than just the global mean warming (Folland
et al. 2001). However, while our knowledge of the connection between human activities
and global mean warming is robust, much uncertainty remains in our understanding of how
some other aspects of the observed climate change relate to anthropogenic forcing.

1.2 Global Warming

The surface warming during the past century is well established. It has been observed both
over land, with station measurements (Jones et al. 2001), and independently over the ocean,
through shipboard measurements (Jones et al. 2001). Large scale surveys of proxy indicators
such as borehole temperature profiles (Beltrami 2002), glacial extent (Folland et al. 2001)
and tree rings and ice cores (Mann et al. 1999), corroborate the warming. Shipboard and
buoy measurements find a corresponding warming in the ocean (Levitus et al. 2000), while satellite measurements over the past three decades indicate that it may extend through the lower troposphere (Prabhakara et al. 2000, Folland et al. 2001), although the satellite measurements are subject to sampling issues that are not yet fully resolved.

During this warming period atmospheric concentrations of greenhouse gases have also been rising. The most important of these is CO$_2$, of which there is now 30% more than before the industrial revolution (Etheridge et al. 1996) and probably more than at any point during at least the past 420 000 years (Petit et al. 1999). The greenhouse gases methane (CH$_4$), nitrous oxide (N$_2$O), tropospheric ozone (O$_3$), and halocarbons (e.g. HFCs, PFCs, CFCs) have also increased during this time (Prather et al. 2001). All of these increases are attributed to human activities; for instance, the burning of fossil fuels is responsible for the rise in CO$_2$ and N$_2$O. In fact most of the halocarbons are not produced naturally (Butler et al. 1999).

Radiative theory dictates that such an increase in these greenhouse gases will force an increase in surface temperature. This will be amplified by the positive feedback of the temperature dependence of the atmospheric concentration of water vapour, the most important greenhouse gas. But the climate system is highly complex and nonlinear, with internal dynamics playing an integral role in the radiative interaction with its surroundings. Thus the magnitude of the warming response could differ substantially from that expected by this simple radiative argument. Moreover, human activities are expected to affect the climate in other ways (Ramaswamy et al. 2001), for instance by producing aerosols (Penner et al. 2001). These aerosols scatter incoming sunlight back into space and also affects the lifetime and optical properties of clouds. Furthermore, external natural forcings, such as changes in solar intensity and the ejection of dust and aerosols from volcanic eruptions also have potentially important effects on the planet’s radiation budget (Ramaswamy et al. 2001).

Consequently, experiments have been conducted to quantify the importance of each of
these effects. Detailed numerical models have been constructed, representing the thermodynamics, moisture transfers, and dynamics of the atmosphere, ocean, sea ice, and land surface systems (McAvaney et al. 2001). Simulations with these models have been conducted which include representations of the evolution of some of these forcings since the industrial revolution (Cubasch et al. 2001). These simulations suggest that only the effects of human activities can reproduce the observed accelerated warming of the past thirty years, although natural forcings were possibly important before then (Stott et al. 2000, Mitchell et al. 2001).

1.3 Changes in Other Climate Properties

The observed climate change has not been limited to global mean temperature. For example, measurements of both precipitation and cloud cover indicate an increase during the past century (Folland et al. 2001). Analyses of the precipitation change suggest that it has resulted from an increase in the frequency of heavy events (Folland et al. 2001). Atmospheric circulation has also shifted systematically, for instance with the mid latitude jets in both hemispheres tending toward more poleward positions (Thompson et al. 2000, Thompson and Solomon 2002). Moreover, the surface has warmed through a detailed spatial and temporal pattern. For example, most of the warming has occurred during the nighttime, with daytime temperatures rising at only half the rate (Easterling et al. 1997, Jin and Dickinson 2002).

Many of these changes are predicted by the numerical models to occur under enhanced greenhouse forcing (Cubasch et al. 2001). However, these predictions have been compared to the observations much less rigorously than is the case with the mean warming. Consequently, it not yet clear with many of these properties whether the magnitude and spatio-temporal patterns of the observed trends are consistent with anthropogenically forced climate change. Nor is it clear in some cases whether such changes are outside of the variability expected through natural fluctuations internal to the climate system, for example in the
case of Northern Hemisphere extratropical circulation (Cubasch et al. 2001). Nevertheless, changes in some of these climate properties could more strongly impact human society than will the mean warming.

This study aims to investigate the changes in two atmospheric properties in global warming simulations of state-of-the-art climate models in order to refine our knowledge of the anthropogenic component in the observed changes. The first of these is the diurnal temperature range, the difference between the daytime maximum and nighttime minimum temperatures. The second is tropospheric circulation, and in particular how changes may manifest themselves as shifts in pre-existing modes of natural variability.

1.4 Diurnal Temperature Range

The mean surface air temperature over land areas increased at a rate of 0.9°C per century over the 1950-1993 period (Folland et al. 2001). Analysis of station measurements indicates that the observed warming resulted largely from a general rise in daily minimum temperatures \( T_{\text{min}} \), with the increase in the daily maximum \( T_{\text{max}} \) being only about half as large (Easterling et al. 1997). The trend in the diurnal temperature range (DTR, the difference between \( T_{\text{max}} \) and \( T_{\text{min}} \)) amounts to \(-0.8\)°C per century, comparable to the mean warming itself. This differential temperature trend is a distinct characteristic of recent climate change, and thus could serve as a “fingerprint” for the identification of the natural and anthropogenic causes of the overall warming. However, despite the magnitude of this trend, its underlying cause and its relation to anthropogenic emissions remain poorly understood.

A first step in improving our knowledge of the observed DTR changes is to determine whether they are consistent with climate model simulations of enhanced greenhouse warming. Consequently, Chapter 2 consists of a direct comparison of model simulations with observed changes in the DTR. A preliminary examination of possible mechanisms linking the changes to the anthropogenic forcing is also undertaken. Chapter 3 contains a detailed
examination of the seasonal and spatial patterns of the model DTR changes through 2100. Following from Chapter 2, a more detailed examination of the robustness and physical basis of the links to anthropogenic forcing is then performed in Chapter 3.

1.5 *Tropospheric Circulation*

Much of the variation in the climate at Earth’s surface can be described as fluctuations in the amplitude of large scale anomaly patterns (e.g. Wallace and Gutzler 1981, Barnston and Livezey 1987). Recently there has been considerable interest in the possibility that anthropogenically forced climate change may manifest itself by projecting directly onto these pre-existing natural modes of variability (e.g. Thompson and Wallace 1998, Fyfe *et al.* 1999, Palmer 1999). In other words, much of the change could adopt the spatial form of these patterns, and thus be represented as simple shifts in measures of these modes. This possibility, as well as the manner in which these changes could occur, is examined in Chapter 4 using simulations made with a global climate model.

The most dominant mode of variability in tropospheric circulation in the Southern Hemisphere represents north-south fluctuations in the position of the mid latitude jet, and is known as the Southern Annular Mode (SAM, also known as the Antarctic Oscillation) (Rogers and van Loon 1982, Nigam 1990, Thompson and Wallace 2000). A gradual poleward shift of the jet, similar to the positive phase of the SAM, is noted in Chapter 4 and has also both been observed (Thompson *et al.* 2000, Thompson and Solomon 2002) and noted in global warming simulations of climate models (Fyfe *et al.* 1999, Kushner *et al.* 2001). Chapter 5 consists of an examination of how this response depends on the surface warming resulting from the use of different schemes for the parametrisation of sub-grid scale ocean mixing.
1.6 Outline

The aim of this investigation is to improve our understanding of how observed changes in two properties of the climate system relate to anthropogenically forced warming. Chapters 2 and 3 contain an examination of the DTR changes in model simulations and in observations, while Chapters 4 and 5 consist of a similar study of changes in tropospheric circulation. The studies in Chapters 2, 3, and 5 all use simulations of two versions of the Canadian Centre for Climate Modelling and Analysis global climate model, and these are described in detail in Section 2.2. On the other hand, the study in Chapter 4 uses simulations of the Geophysical Fluid Dynamics Laboratory global climate model, which are described in Section 4.2. Finally, Chapter 6 consists of a discussion of the results and their implications.
Chapter 2

Daily Maximum and Minimum Temperature Trends

2.1 Introduction

The observed global mean trend towards warmer temperatures over land has been characterised by a large increase in the minimum daily temperatures ($T_{min}$) (Karl et al. 1993, Easterling et al. 1997, New et al. 2000, Jin and Dickinson 2002). Maximum daily temperatures ($T_{max}$) have increased at a much smaller rate, resulting in a decreasing trend in the diurnal temperature range (DTR) over land, the magnitude of which is comparable to the mean warming itself. As an identifiable characteristic of recent climate change, this trend is important in diagnosing the forcing responsible for the change, and in particular the anthropogenic component. However, the cause of the DTR trend is still poorly understood, as is its relation to anthropogenic forcing.

Observational studies suggest that regions where the DTR has decreased have also tended to experience an increase in low base clouds (Karl et al. 1993, Dai et al. 1997b, 1999). The reflection of sunlight by these low level clouds would be expected to cause a drop in daytime temperatures, and indeed modelling studies indicate that the DTR would be quite sensitive to changes in cloud cover (Stenchikov and Robock 1995, Dai et al. 2001). Thus increasing cloud coverage is suggested as the primary cause of the observed DTR decrease. However, increases in soil moisture, by controlling evaporation and the ground
heat capacity, as well as in sulphate aerosols, by scattering sunlight back to space, could also strongly influence the DTR. Moreover, due to difficulties in the availability and accuracy of measurements, observational verification for these relations is difficult, and thus model investigations are required.

Studies with global climate models project a decrease in the DTR under enhanced greenhouse forcing (Cao et al. 1992, Colman et al. 1995, Mitchell et al. 1995, Stenchikov and Robock 1995, Reader and Boer 1998, Dai et al. 2001). These investigations indicate that the DTR is relatively insensitive to short wave scattering by sulphate aerosols, but is influenced by clouds and soil moisture, as well as plant physiological responses (Collatz et al. 2000). In this chapter, we evaluate the observed DTR trend as a potential response to anthropogenic forcing by directly comparing simulations of a global climate model with observations, and relate the DTR trend to trends in other variables describing the climate system.

2.2 Methods

This investigation uses simulations from the first generation coupled general circulation model of the Canadian Centre for Climate Modelling and Analysis (CCCma), known as CGCM1 (Flato et al. 2000). It includes comprehensive representations of the atmosphere, ocean, sea ice, and land surface. Reader and Boer (1998) also investigate DTR trends in simulations of a slab ocean version of this model, but here we conduct a direct comparison with observed changes.

The atmospheric component of CGCM1 is a spectral model with triangular truncation at wavenumber 32 (McFarlane et al. 1992). This yields a surface grid resolution of about 3.75°. A hybrid topographic-pressure vertical coordinate system is employed resulting in 10 unequally spaced levels. A land surface scheme underlies the atmospheric component, and uses a single soil layer with spatially varying moisture field capacity and soil properties.
Screen level temperature is estimated from temperatures at the lowest level (200 m) and the surface using a gradient profile relationship. To adequately resolve the diurnal cycle, full solar radiation calculations are performed every three hours, with the radiation at intervening time steps (20 minutes) extrapolated from the full calculation based upon the solar zenith angle. Full calculations of the terrestrial radiation are performed every six hours, with partial calculations at intervening time steps which re-evaluate the fluxes and heating rates using emissivities computed during the previous full calculation.

The ocean component is a grid point model with double the horizontal resolution of the atmosphere (1.875°) and 29 unequally spaced vertical levels. A simple one dimensional thermodynamic sea ice model is used. In order to prevent drift of the model integrations toward less realistic states, heat and fresh water flux adjustments are used between the atmosphere and ocean (Flato et al. 2000).

We use an ensemble of three simulations (CGCM1 GHG+A1,2,3) which include observed increases in greenhouse gases as well as the scattering of sunlight by increases in sulphate aerosols, and projected changes approximately following the IPCC’s IS92a “business as usual” scenario (Boer et al. 2000b). The aerosols are represented through changes in the surface albedo (Reader and Boer 1998); the so-called indirect effect of aerosols, involving changes in the lifetime and optical properties of clouds, is not included. These simulations span the 1850-2100 period and are identical except for their initial conditions, and so represent independent possible realisations of recent climate. The analysis here makes use of output from the 1950-1993 period, which is the period for which observations are available. These model integrations, and other simulations using the same model which are used in later chapters, are listed and compared in Table 2.1.

We also examine an ensemble of three integrations of a more recent version of the CC-Ccmh coupled model, known as CGCM2 (CGCM2 GHG+A1,2,3). This model uses different representations of ocean mixing and sea ice than CGCM1 (Flato and Boer 2001). In particular, the Gent and McWilliams parametrisation associated with mesoscale eddies (Gent
and McWilliams 1990) and a cavitating fluid representation of sea ice (Flato and Hibler 1992) are included. An important result of these modifications is a much larger warming of the surface at middle and high southern latitudes, producing a more meridionally symmetric warming pattern (Flato and Boer 2001). The details of the integration procedure are essentially the same as the CGCM1 GHG+A integrations.

### 2.3 Results

We first compare annual mean trends in the DTR from the model simulations with those from the observations. Values from the model grid are interpolated to the observational grid, and retained only where observational measurements exist (Easterling et al. 1997). This observational grid covers most of the Eurasian, North American, and Australian landmasses at 5° resolution in both latitude and longitude, as well as areas of Africa and South America, and many oceanic islands. The global model trend in \( T_{\text{max}} \) ranges from 1.3 to 1.5°C per century (Figure 2.1). We use a 201 year simulation with constant forcing (CGCM1 CTRL) to estimate the range of 44 year trends plausibly due to natural variability by taking eight overlapping 44 year segments. With this estimate of the natural variability, we find that the trend in annual mean \( T_{\text{max}} \) in the model simulations is significantly higher than the 0.8°C per century inferred from observational measurements (at the 5% level for a two-sided test).
On the other hand, the global increases in $T_{\text{min}}$, ranging from 1.5 to 1.7°C per century, are consistent with the observed warming of 1.8°C per century. In both cases the trends are significantly different from zero. As with the observations, these differential temperature trends in the model integrations result in a significant decrease in the DTR (Figure 2.2). However, this change of -0.2°C per century is considerably, and significantly, smaller than the observed -0.8°C per century.

In the Northern Hemisphere, decreases occur in the DTR in all seasons in the model simulations, with those during autumn, winter, and spring being statistically significant (Figure 2.1). The modelled trends tend to be largest in the winter, as is the case in the observations. In all seasons, however, the model underestimates the observed trends. During winter and spring the modelled and observed trends in $T_{\text{max}}$ agree well with each other. The increases in $T_{\text{min}}$, on the other hand, tend to be smaller in the model, and this results in the smaller decrease in the DTR. During the summer and autumn, on the other hand, $T_{\text{min}}$ trends are better reproduced in the simulations than are $T_{\text{max}}$ trends, such that the DTR trends are underestimated in the model due mainly to an overestimate of the warming in $T_{\text{max}}$. Possible reasons for this seasonal pattern are discussed below.

Unlike in the Northern Hemisphere, the seasonal and annual DTR trends in the Southern Hemisphere in the model simulations are not significantly different from zero. With the exception of winter, the decreases are considerably smaller than the observed values. As is the case for the Northern Hemisphere, the underestimation results more from an overestimation of the $T_{\text{max}}$ warming during the summer and autumn, and an underestimation of the $T_{\text{min}}$ warming during the spring. The large spread of trend estimates between the model simulations indicates that the size of the station network in the Southern Hemisphere is as yet insufficient to robustly estimate the DTR trends. In fact, trends calculated for all Southern Hemisphere land areas (excluding Antarctica) are systematically less negative in the simulations. However, this sampling effect of about ±0.3°C per century is not large enough to affect the sign of the observed -0.6°C per century decrease.
Figure 2.1: 1950-1993 seasonal trends in $T_{\text{max}}$, $T_{\text{min}}$, and the DTR. Global and hemispheric trends are shown. The red values are the observed trends at nonurban stations in Easterling et al. (1997); values from the three CGCM1 GHG+A model simulations are in dark blue, while those from the CGCM1 GHG simulation are in light blue. The values in green are from simulations including both greenhouse gases and sulphate aerosols using a newer version of the CCCma model (CGCM2 GHG+A1,2,3). The error bars denote the 95% confidence interval of the CGCM1 GHG+A simulations about the mean of the CGCM1 GHG+A results, calculated from the natural variability of the CGCM1 CTRL simulation. DJF is the December-February season, MAM is March-May, JJA is June-August, and SON is September-November.
Figure 2.2: 1950-1993 time series of annual mean DTR. The time series from nonurban station measurements, from Easterling et al. (1997), is in red, while the blue lines represent the time series from the three CGCM1 GHG+A model simulations. Values are anomalies from the 1950-1959 mean.

The global warming simulations of CGCM2 predict $T_{max}$, $T_{min}$, and DTR trends in the Northern Hemisphere similar to those in the GHG+A simulations of CGCM1 (Figure 2.1). Since most of the land in this hemisphere is far away from the ocean, it is not surprising that a different ocean component for the model has no obvious effect on the DTR over land. However, ocean dominates over land in the Southern Hemisphere, so most land is in relative close proximity to the ocean. Indeed, the DTR tends to decrease more in these simulations than in those from CGCM1, and they more closely resemble the observations. This better agreement arises from improved prediction of both the $T_{max}$ and $T_{min}$ trends, and demonstrates that better representation of high latitude ocean and sea ice processes is necessary to properly represent Southern Hemisphere climate, even over land areas.

Observed changes in the DTR are not uniform. For instance, increases have actually occurred over parts of Canada and the South Pacific islands (Figure 2.3). Similar regional increases also occur in the model simulations, but the patterns do not resemble those that are observed. However, the regional changes in the different simulations are also rather different from one another. For instance, in one simulation the DTR decreases uniformly
over Australia, while in another it increases and in the third remains fairly constant. This indicates that regional trends in the DTR may not be distinguishable from random natural variability over the rather short 1950-1993 period.

The most notable difference between the spatial pattern of the DTR trend between the model simulations and the observations is over island areas, especially in the South Pacific Ocean. This discrepancy arises from differences in the representation of these areas. Observational measurements come from land stations, and thus are biased toward the small islands. On the other hand, since there is more water than land in these grid boxes they are represented as ocean in the model. The DTR over the sea is considerably smaller than over land due to the large thermal capacity of water. Consequently, long term trends would also be smaller over water than on land, and most likely not even detectable at this stage. However, removal of these island areas from the comparison amounts to only a further 0.05°C per century decrease in the DTR.

The scattering of sunlight by increasing concentrations of sulphate aerosols could be a cause of the DTR decrease, since this diminishes the amount of energy reaching the surface during the daytime. However, results from a simulation of the model omitting increases in aerosol concentrations, but still including those in greenhouse gases (CGCM1 GHG), indicate that this is not the case (Figure 2.1). While the warming of $T_{min}$ and $T_{max}$ is much larger in this CGCM1 GHG simulation, changes in the DTR are similar to those in the CGCM1 GHG+A simulations in most seasons. Thus in the model simulations the DTR decrease is a result of the increase in greenhouse gases and is largely independent of the emission of sulphate aerosols, as found in other spatial-temporal domains (Reader and Boer 1998) and other models (Cao et al. 1992, Mitchell et al. 1995, Stenchikov and Robock 1995). Stenchikov and Robock (1995) suggest that the effect of aerosol scattering on the DTR is cancelled by a water vapour feedback. In the cooler climate resulting from the aerosols, less atmospheric water vapour is present to absorb near-infrared solar radiation, thus increasing the total radiation reaching the surface.
Figure 2.3: 1950-1993 DTR trends in the observations and the three CGCM1 GHG+A model simulations for each 5° by 5° grid box. The observations are from Easterling et al. (1997) and are calculated from nonurban stations. The scale is identical for all maps.
Other suggested influences on the DTR trend are increases in cloud cover and soil moisture. During the daytime clouds reduce the amount of sunlight reaching Earth's surface. Increases in soil moisture permit faster cooling during the daytime through evaporation and also moderate temperatures by increasing the heat capacity of the ground. Our analysis suggests that changes in these two factors are indeed related to the DTR trend in the model simulations. We create a least squares estimated multiple regression model of the effects of changes in annual mean daytime cloud cover and in soil moisture on the DTR:

$$\Delta T_G \sim \Delta T_C + (f_G + \bar{f}_G) \frac{s_{\Delta T_C}}{s_{f_C}} \left( \frac{1}{1 - r_{f,\Delta T_C}} \right) \left( r_{f,\Delta T_C} - r_{w,\Delta T_C} \right)$$

$$+ (w_G + \bar{w}_G) \frac{s_{\Delta T_C}}{s_{w_C}} \left( \frac{1}{1 - r_{f,\Delta T_C}} \right) \left( r_{w,\Delta T_C} - r_{f,\Delta T_C} \right)$$

(2.1)

Here $\Delta T$ is the DTR, $f$ is the cloud cover, and $w$ is the soil moisture. $s_x$ denotes the estimated standard deviation of variable $x$, while $r_{x,y}$ denotes the estimated correlation between variables $x$ and $y$. The $C$ subscript indicates values derived from the 201 years of the CGCM1 CTRL simulation, while the $G$ subscript indicates values from the CGCM1 GHG+Al simulation. $\Delta T_G$ is then the estimated DTR for the CGCM1 GHG+Al simulation. Daytime cloud cover is measured here by the amount of solar radiation reaching the ground. The correlation between the estimated and actual DTR time variations for the CGCM1 GHG+Al simulation is 0.91 over the 1950-2100 period, while their trends are both -0.22°C per century (Figure 2.4). Use of only one of these variables (solar radiation or soil moisture) to predict the DTR results in considerably less accurate correspondences, indicating that both are important. Previous modelling studies (Stenchikov and Robock 1995, Collatz et al. 2000, Dai et al. 2001) support this result. Interestingly, variations and changes in the mean temperature and DTR are rather unrelated (not shown).

The influence of soil moisture on the DTR arises through a number of mechanisms. Variations in soil moisture are caused by changes in evaporation and precipitation, both of which are related to cloud cover. Thus the relations between soil moisture and the
Figure 2.4: Time series of annual mean DTR from 1950 through 2100 in the GHG+A1 simulation. The solid line is the actual value from the simulation, while the dotted line is estimated using a regression model from variations in daytime cloud cover and in soil moisture.

DTR could simply reflect the correlation of both to cloud cover. However, removal of the component of the soil moisture variations correlated with cloud cover reveals that the residual still has an important relation to the DTR variations and trend. Another possibility is that soil moisture influences the DTR through changes in evaporation and vegetative evapotranspiration (Collatz et al. 2000). However, in the model simulations this effect is largely cancelled by opposite changes in the sensible heat flux (not shown). The possibility that soil moisture acts as a proxy for the water vapour radiative feedback described by Stenchikov and Robock (1995) is not supported because of the lack of covariation between specific humidity and the DTR (not shown). This leaves changes in the moderating effect of the heat capacity of the ground as the main mechanism relating soil moisture to the DTR decrease in the model. Of course, the importance of this mechanism may be magnified by the use in CGCM1 of the single layer bucket model in representing the land surface.

The reflection of incoming solar radiation by cloud cover serves to reduce $T_{\text{max}}$, leaving $T_{\text{min}}$ relatively unaffected. Thus underestimates of daytime increases in cloud cover in the CGCM1 GHG+A1 simulation would result in the overestimate of $T_{\text{max}}$ during summer, as noted earlier. During this season the amplitude of the diurnal cycle of solar radiation is
highest, and so $T_{max}$ would be most sensitive to the reflection of sunlight by clouds. During the winter, on the other hand, the presence of snow both reflects much of the incoming sunlight and insulates the atmosphere from the soil moisture. The importance of clouds during this season lies in their downward emission of infrared radiation, which serves to maintain temperatures overnight. Therefore, an underestimate of increasing cloud cover during the winter would result in an underestimate of the $T_{min}$ trend, such as occurs in the CGCM1 GHG+A simulations. The importance of cloud cover to the DTR is evidenced by the CGCM2 GHG+A simulations. These and the CGCM1 GHG+A simulations differ mainly in that the former predict a much larger warming of the surface ocean and atmosphere in the Southern Hemisphere as a consequence of an improved representation of ocean mixing. A warmer atmosphere over a warmer ocean implies more moisture in the air, which then forms clouds when passing over land. Indeed, cloud cover increases substantially in the Southern Hemisphere in the CGCM2 GHG+A simulations, producing better agreement with the observed temperature trends. Therefore, an underestimate of an increase in global cloud cover over land in the model simulations could account for much of the discrepancy between the modelled and observed trends in the DTR.

Forced with predicted changes in greenhouse gases and sulphate aerosols, CCCma models project a continued decrease in the DTR through the twenty-first century (Figure 2.4). During the December-February and March-May seasons the DTR is projected to decrease at a similar rate as in the 1950-1993 period. On the other hand, little change is projected to occur during the June-August and September-November seasons. Due to the spatial bias of the stations, this pattern of change is dominated by large decreases in the Northern Hemisphere during the winter and spring.
2.4 Conclusion

These results indicate that the observed decreases in the DTR could be a climatic response to anthropogenic emissions of greenhouse gases and aerosols. In particular, changes in cloud cover and soil moisture associated with climate change force the DTR reduction in models. However, a discrepancy in the magnitude of the trend between observations and the model simulations remains. The importance of soil moisture found here implies that physiological responses of vegetation to climate change could be quite important for the behaviour of the DTR. Improvements in the parametrisation of clouds and land surface processes are currently among the most actively pursued goals in climate model development. Thus more reliable estimates of the importance of the observed DTR trend as a fingerprint of anthropogenic forcing of climate change can be expected in the near future. At this early stage, however, model results are consistent with the observed DTR decrease over the last half century, and suggest that this trend is likely to continue into the foreseeable future.
Chapter 3

Factors Contributing to DTR Trends in Model Simulations

3.1 Introduction

In the last Chapter, we found that a decrease in the DTR is predicted by the CCCma models to occur under enhanced greenhouse forcing. We also noted that this change coincides with trends in two other facets of the climate system, namely clouds and soil moisture. However, the robustness of these links remain unclear, as does their physical cause. Consequently, in this chapter we conduct a more thorough investigation of the causes of the DTR decrease in the model integrations.

Variations in cloud cover are strongly correlated with those in the DTR (Dai et al. 1997a, 1999, New et al. 2000). The higher albedo of clouds decreases the downward solar radiation during the day, and thereby reduces $T_{max}$. Indeed, observational studies link the decreasing DTR to coincident increases in precipitating clouds (Karl et al. 1993, Dai et al. 1997a, 1999). These low base clouds are particularly effective in reflecting sunlight, and changes in their frequency of occurrence would be expected to have the strongest impact on the DTR. Clouds also emit more downward long wave radiation, so increasing nighttime cloud cover would increase $T_{min}$ and thereby decrease the DTR. However, the tendency of the diurnal cycle of cloud cover over global land areas during recent years is currently unknown.
Soil moisture is also expected to influence the DTR, through control of evaporative cooling, the ground albedo, and the ground heat capacity. This effect tends to be most influential in the occurrence of extreme hot days (Durre et al. 2000). Dai et al. (1999) find that soil moisture is related to the DTR, albeit secondary to changes in cloud cover. This raises the possibility that the observed decreases in the DTR may be due in part to physiological responses of vegetation to climate change or to changes in land use, although it appears that this latter factor is unable to fully account for the observed changes (Easterling et al. 1997, Gallo et al. 1999).

Atmospheric and coupled general circulations models (GCMs) predict a decrease in the DTR under enhanced greenhouse forcing (Cao et al. 1992, Mitchell et al. 1995, Colman et al. 1995, Reader and Boer 1998, Dai et al. 2001), but it was found in Chapter 2 that the magnitude of this change from at least one model is considerably smaller than observed. In agreement with energy balance models (Cao et al. 1992, Stenchikov and Robock 1995), the addition of the scattering effect of sulphate aerosols produces little difference in the DTR change (Mitchell et al. 1995, Reader and Boer 1998). Stenchikov and Robock (1995) note that in an energy balance model the DTR is quite sensitive not only to mean changes in cloud coverage, but also to the nature of the diurnal cycle of the coverage. Indeed, Dai et al. (2001) find in a GCM integration that the reduction in the DTR is associated with changes in cloud coverage, as well as with changes in soil moisture. However, Collatz et al. (2000) note that the physiological response of vegetation, which is not represented in these models, could also be a rather important influence.

To better understand its relation to current climate change this chapter consists of an examination of the nature and cause of trends in the DTR in integrations of a coupled GCM. The model and the integrations were described in Section 2.2, with a brief elaboration of some relevant details given in Section 3.2. Section 3.3 consists of a detailed examination of the DTR trends produced in the model integrations. In Section 3.4 a simple analytic model is used to diagnose possible causes for the trends, while in Section 3.5 statistical models are
used to further isolate these causes. The cause of the DTR trends in the middle latitude winter is not evident from the analyses in these two sections, and thus is examined more closely in Section 3.6. The results are discussed in Section 3.7.

3.2 Model

The models used in this investigation are CGCM1 and CGCM2, which are described in detail in Section 2.2. We use the same three global warming simulations of each as in Chapter 2, although we concentrate on the output of CGCM1 GHG+Al. In these integrations the model is forced with the observed atmospheric concentrations of greenhouse gases and sulphate aerosols until present, and with those projected for the future according to a modified version of the IPCC 1992a scenario (Boer et al. 2000a,b). The direct scattering of sunlight by sulphate aerosols is included by altering the surface albedo (Reader and Boer 1998). A similar integration which lacks the scattering effect of sulphate aerosols (CGCM1 GHG) is also examined in order to determine the importance of the aerosols for the DTR trend. Finally, a 201 year control integration of each model version (CGCM1 CTRL and CGCM2 CTRL), which uses constant pre-industrial forcings, is used as a reference. These model integrations are listed and compared in Table 2.1.

Since the climate does not change substantially until 1950, we examine the 1950-2100 interval here. The DTR over the ocean is quite small, and so trends in the DTR are negligible, if even detectable in the observations. Therefore, only land areas are included in the analysis. Areas poleward of the Arctic and Antarctic circles are also excluded since the DTR does not represent the daytime-nighttime cycle at these latitudes. In the CGCM1 simulations, the southern tip of Greenland (which lies south of the Arctic circle) is permanently covered in snow. Since this complicates the comparison of the DTR to other climate variables, this region is also excluded from the analysis.
3.3 Trends

In Chapter 2 we directly compared the DTR trends in CGCM1 integrations with those observed by Easterling et al. (1997) over the 1950-1993 period. While the integrations indicate a tendency toward decreasing DTR, these trends are smaller than observed by about two thirds. Here we more closely examine the spatial nature of the model trends over land areas, looking particularly at the 1950-2100 period. The annual mean time series of the DTR, $T_{\text{min}}$, and $T_{\text{max}}$ in CGCM1 GHG+A1 over this interval are shown in Figure 3.1. The data used cover all land areas (except for the polar restrictions). The DTR decrease, which starts around 1970 and is rather linear thereafter, is much smaller than the trend in the mean temperature.

The mean annual trends in the DTR during the December-January (DJF) and June-August (JJA) seasons over the 1950-2100 period from CGCM1 GHG+A1 are displayed in Figure 3.2. The spatial pattern is very similar in the other global warming integrations. It does not, however, resemble the observed spatial pattern of Easterling et al. (1997), though in Chapter 2 we found that the observed pattern may not be robust over the 44 years of
Figure 3.2: 1950-2100 mean trends in the DTR in CGCM1 GHG+A1 during the DJF and JJA seasons. Values are shown over non-polar land areas only. Solid bullets denote trends statistically significant at the 5% level, assuming a white noise process.

observations. The DTR rather uniformly decreases in the middle northern latitudes, but in other areas there is a more regional mix of positive and negative trends. Most of the differences occur on regional scales, with the pattern being rather smooth at the smaller scales near the model resolution.

To better understand this pattern of trends, we examine them separately for each season and for four zonal bands, covering the 66°S-33°S, 33°S-0°, 0°-33°N, and 33°N-66°N intervals. The borders between these bands are indicated in Figure 3.2 by the dashed lines. These
**Figure 3.3:** Annual and seasonal mean trends in the DTR during the 1950-2100 interval over four zonal bands. Values are calculated for land areas only. For each season the three blue bars represent trends in the CGCM1 GHG+ A integrations, while the cyan bar represents the trend in CGCM1 GHG. The green bars represent trends in three integrations of the CGCM2 model forced with changes in greenhouse gases and sulphate aerosols similar to those imposed in the CGCM1 GHG + A integrations.

Divisions correspond approximately to changes in the seasonal behaviour of the DTR and in its relation to other climate variables. In particular, the relative importance of $T_{\text{min}}$ and $T_{\text{max}}$ in determining the DTR changes near 33° of latitude during the winter (and near 66° of latitude during the summer) in both hemispheres (not shown). Trends over these regions in the CGCM1 GHG + A integrations during the 1950-2100 period are displayed in Figure 3.3.

Some of the largest DTR decreases occur in the middle northern latitudes during DJF and March-May (MAM), while changes are minimal in the other two seasons. The mean temperature rises faster in DJF and MAM, and so the DTR decreases result from particularly large increases in $T_{\text{min}}$ (not shown). Decreases in the DTR about half as large
also occur in the low northern latitudes in DJF and MAM, as well as during September-November (SON). The mean warming is also very similar in these seasons. The exception of JJA results mainly from a slower warming of $T_{\text{min}}$.

DTR changes are small in all seasons in the low southern latitudes, with mean temperature increases also rather similar in each season. However, the middle southern latitude DTR trends are of comparable magnitude to those in the middle northern latitudes, while the mean warming is much smaller than in the other regions. This region experiences large decreases in the DTR in SON, DJF, and MAM. The anomalous increase in the DTR during JJA is due to a smaller warming of $T_{\text{min}}$. The similarity to the calendar season pattern in northern middle latitudes is puzzling considering the phase difference in the seasons between both hemispheres. Of course, while the trends appear consistent between integrations, it should be remembered that the southern middle latitudes contain very little land. Furthermore, this land is close to the low latitudes, and so the climatology here may not properly represent that over middle latitude land masses.

In Chapter 2 we noted that the CGCM2 GHG+A integrations predict more negative DTR trends in the Southern Hemisphere than do the GHG+A integrations of CGCM1, in better agreement with the observations, whereas the two models differ little in the Northern Hemisphere trends. The important difference between the two models is the use of the Gent and McWilliams mixing parametrisation in CGCM2, which results in a larger warming in the Southern Hemisphere. The hemispheric characteristics of the DTR trends from the two models also hold over the longer 1950-2100 period and full global landmass (Figure 3.3). Practically identical trends occur over the Northern Hemisphere in the integrations of both models during most seasons. In the lower southern latitudes, however, the trends tend to be more negative in the CGCM2 GHG+A integrations, although in both models the change is small. Over the middle southern latitudes, on the other hand, the trends are substantially more positive. However, since most of the Southern Hemisphere landmass is in the lower latitudes, the hemispheric trends closely resemble those for this region.
from Chapter 2 suggest that this difference between the model results in the Southern Hemisphere arises from increased clouds over land produced by the warmer ocean. Indeed, the decrease in solar radiation at the surface is 40% larger over the Southern Hemisphere in the CGCM2 GHG+A integrations than in those of CGCM1, whereas little difference exists in the Northern Hemisphere (not shown). The importance of clouds for the DTR trend is examined further in the next two sections.

Trends in the DTR in CGCM1 GHG are also displayed in Figure 3.3. The single difference between this and the CGCM1 GHG+A integrations is the absence of scattering due to sulphate aerosols. This results in a considerably larger mean warming. However, as noted previously (Mitchell et al. 1995, Reader and Boer 1998), exclusion of this forcing produces little change in the DTR trend. In some cases the omission results in very slightly more positive trends, but the general effect is minimal.

3.4 Physical Analysis of Factors Influencing the DTR

The diurnal temperature range is influenced by several factors, many of which could change under global warming. To diagnose the relative importance of these factors we formulate a first order analytic calculation of the DTR. The intention here is simply to identify climate variables to which the DTR is directly most sensitive, and which therefore could induce a trend under global warming. Of course, this calculation will require many approximations, but the advantage lies in the simplicity of its physical interpretation. We stress that this calculation is intended for interpretive purposes only, and is not intended to exhaustively represent all of the processes operating during the diurnal cycle. More accurate, but less physically intuitive, statistical models will be used in the next section to support the conclusions from this analytic calculation.
The rate of change of the temperature of the surface is given by

\[
\frac{\partial T}{\partial t} = S + L_\updownarrow + L_\uparrow + H_l + H_s + U. \tag{3.1}
\]

Here \(c_a\) is the effective heat capacity of the ground, \(T\) is the surface temperature, \(t\) is time, \(S\) is the incoming solar radiative flux, \(L_\downarrow\) and \(L_\uparrow\) are the incoming and outgoing terrestrial long wave radiative fluxes, \(H_l\) and \(H_s\) are the latent and sensible heat fluxes, and \(U\) is the horizontal advective term. We assume that the surface and screen level temperatures are identical, and so use the latter since it is archived from the model simulations. The heat capacity of the air near the surface is considerably smaller (about two orders of magnitude) than that of the ground, so only the ground heat capacity is included here.

We start by taking the diurnal cycle to be a step function, jumping instantaneously from nighttime to daytime values. A more accurate description would result in a much more complicated solution than the first order model desired here. Once again, we stress that this calculation is for interpretative purposes only, and thus such approximations are permitted. From Eq. (3.1), the change of surface temperature (\(\Delta T\)) from night to day (\(\Delta t\)) is then

\[
\frac{\Delta T}{\Delta t} \sim S_d + L_{d\downarrow} + L_{d\uparrow} + H_{d\uparrow} + H_{d\downarrow} + U_d
\]

\[
- S_n - L_{n\downarrow} - L_{n\uparrow} - H_{n\uparrow} - H_{n\downarrow} - U_n
\]

(3.2)

where the \(d\) and \(n\) subscripts denote typical daytime and nighttime values.

The daytime solar flux at the surface can be approximated as

\[
S_d \sim S_0(1 - f_c \alpha_c)(1 - \alpha_g)
\]

where \(S_0\) is the average daytime solar radiative flux at the top of the atmosphere, \(\alpha_c\) and \(\alpha_g\) are the cloud and ground albedos, and \(f_c\) is the fraction of the sky covered by cloud.
Since

\[ S_c \sim S_0(1 - f_c \alpha_c) \]  

(3.3)

is archived in the model, while \( \alpha_c \) is not, we use \( S_c \) instead. \( S_n \) is obviously zero.

The incoming long wave flux possesses only a small diurnal cycle (e.g. Dai et al. 1999) which can be considered negligible for the present purposes (\( L_{ld} \sim L_{ln} \)). Meanwhile, under the step function assumption the outgoing long wave radiation is given by \( L_{td} \sim -\sigma T_{max}^4 \) and \( L_{tn} \sim -\sigma T_{min}^4 \), where \( \sigma \) is the Stefan-Boltzmann constant. The emissivity is omitted here since it is prescribed as unity for all surfaces in CGCM1 (Norm McFarlane, personal communication). Upon linearisation about some \( T_0 \) near \( T_{max} \) and \( T_{min} \), this yields a difference of

\[ L_{td} - L_{tn} \sim -4\sigma T_0^3 \Delta T. \]

We combine the latent and sensible heat fluxes since they are forced by similar processes, exhibit similar diurnal cycles (Barry and Chorley 1992), and have identical effects on the screen temperature. We assume zero flux during the night, and a constant value during the day, yielding

\[ H_d = H_{ld} + H_{sd} \]

\[ \sim -H_0. \]

Here \( H_0 \) is taken to be the heat flux averaged over the full diurnal cycle and we assume it to be the averaged daytime value only. This assumption likely overestimates the actual amplitude of the diurnal cycle of the heat fluxes as they are generally not quite zero at night.

The magnitude of the diurnal temperature wave into the ground decreases as a function
of depth, and so using just the heat capacity of soil would be inappropriate. This effect of vertical diffusion is absorbed into the value of $c_g$. Since $c_g$ depends upon the soil moisture, it can be described by

$$c_g \sim c_{g0} + c_{g1} w_g$$

where $c_{g0}$ is the effective heat capacity of dry soil and $c_{g1}$ is the rate of change of the effective heat capacity as a function of the soil moisture fraction $w_g$.

Finally, since advection should be uncoupled from the diurnal cycle at most locations we define $U_d$ and $U_a$ to be equal, and thus they cancel over the diurnal cycle.

Substituting these approximations into Eq. (3.2), we find that

$$\Delta T \sim \frac{S_c(1 - \alpha_g) - H_0}{c_{g0} + c_{g1} w_g + 4\sigma T_0^2 \Delta t} \Delta t.$$  (3.4)

Annual global mean values of the parameters in Eq. (3.4) are listed in Table 3.1. The domain (global or zonal band) average obtained from CGCM1 CTRL is used for parameters other than $\sigma$ and $\Delta t$. Domain averages are used for $c_{g0}$ and $c_{g1}$ (Daniel Robitaille, personal communication). $\Delta t$ is chosen to be nine hours as this is the typical length of time for the increase from $T_{min}$ to $T_{max}$. Estimation of the DTR with these parameter values yields 15 K, which compares reasonably with the actual value of 11 K from CGCM1 CTRL, considering the many approximations. Note that the actual mean value of 11 K found here is larger than the $\sim$7 K indicated in Figure 2.4 due to differences in the spatial domain used in the calculations.

The temporal correlations between $\Delta T$ from Eq. (3.4) and the model DTR averaged over each of the four zonal bands are listed in Table 3.2 for the DJF and JJA seasons. In each row of the table, the time series of each indicated variable is included in the calculation, with the other parameters being held constant. For the results in the Table, values were
Parameter | Name                                | Value
---|--------------------------------------|------
$S_c$  | Mean daytime solar radiation        | 420 W·m⁻²
$\alpha_g$ | Ground albedo                  | 0.24
$\Delta t$ | Time interval             | $3.2 \times 10^4$ s
$c_{d0}$  | Dry soil heat capacity          | $8.5 \times 10^4$ J·K⁻¹·m⁻²
$c_{d1}$  | Wet soil heat capacity factor    | $7.9 \times 10^4$ J·K⁻¹·m⁻²
$w_g$  | Soil moisture fraction           | 0.45
$H_0$  | Diurnal heat flux                | 180 W·m⁻²
$\sigma$ | Stefan-Boltzmann constant        | $5.7 \times 10^{-8}$ W·K⁻⁴·m⁻²
$T_0$  | Mean surface temperature         | 290 K

**Table 3.1:** Annual mean values for the parameters in Eq. (3.4) averaged over global land areas.

<table>
<thead>
<tr>
<th>Variable</th>
<th>66-33°S</th>
<th>33°S-0°</th>
<th>0°-33°N</th>
<th>33-66°N</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>DJF</td>
<td>JJA</td>
<td>DJF</td>
<td>JJA</td>
</tr>
<tr>
<td>$S_c$ (Solar radiation)</td>
<td>0.8</td>
<td>0.2</td>
<td>0.9</td>
<td>0.8</td>
</tr>
<tr>
<td>$\alpha_g$ (Ground albedo)</td>
<td>-0.9</td>
<td>-0.2</td>
<td>-0.5</td>
<td>-0.6</td>
</tr>
<tr>
<td>$H_0$ (Heat fluxes)</td>
<td>0.2</td>
<td>0.1</td>
<td>0.0</td>
<td>0.7</td>
</tr>
<tr>
<td>$w_g$ (Soil moisture)</td>
<td>0.9</td>
<td>0.7</td>
<td>0.8</td>
<td>0.8</td>
</tr>
<tr>
<td>$T_0$ (Mean temperature)</td>
<td>-0.4</td>
<td>-0.7</td>
<td>-0.6</td>
<td>-0.5</td>
</tr>
<tr>
<td>$S_c, \alpha_g, H_0, w_g, T_0$</td>
<td>1.0</td>
<td>0.4</td>
<td>1.0</td>
<td>0.9</td>
</tr>
<tr>
<td>$S_c, w_g$</td>
<td>0.9</td>
<td>0.5</td>
<td>0.9</td>
<td>0.9</td>
</tr>
</tbody>
</table>

**Table 3.2:** The temporal correlation between the model DTR from CGCM1 CTRL and estimates from Eq. (3.4). For each estimate, the time series from CGCM1 CTRL is used for the indicated variable(s); other variables are assumed constant at their mean value. Seasonal means of the variables were averaged over the zonal bands before input into the equation. Coefficients outside of the -0.1 to 0.1 interval are significant at the 5% level assuming white noise processes.

averaged over space before being input into Eq. (3.4). The correlations at each grid point for annual mean values are displayed in Figure 3.4.

$S_c$ and $w_g$ are both individually highly correlated with the DTR in most cases. Little skill is added by the inclusion of the variability from the other factors. While the sensible and latent heat fluxes are each highly correlated with the DTR (not shown), their sum is mostly unrelated to it, consistent with the results of observational studies (Dai et al. 1999).

Exceptions to the high correlation between both $S_c$ and $w_g$ with the DTR occur during the winter seasons in the middle latitudes. During these seasons snow covers most of these
Figure 3.4: Map of the correlation between the annual mean DTR in CGCM1 CTRL and estimates from Eq. (3.4). Annual mean values for $S_c$, $\alpha_g$, $H_0$, $w_g$, and $T_0$ from CGCM1 CTRL were used in the calculation.

areas, which reflects sunlight and thus reduces the diurnal amplitude of the solar radiative forcing. Also, snow insulates the air from the ground, damping the moderating effect of the soil moisture's heat capacity. It is not surprising that neither variable is strongly related to the DTR in these cases. Unfortunately, it is not clear from Table 3.2 what replaces these factors in controlling the DTR; this will be examined more closely in Section 3.6.

The 1950-2100 DTR trends in CGCM1 GHG+A1 predicted by Eq. (3.4) with the inclusion of the changes in the various variables are compared to the model DTR trends in Table 3.3. The DTR tendency is most sensitive to the changes in $S_c$, but the estimated trends are considerably more negative than the actual model DTR trends in most cases. Inclusion of the changes in the remaining variables does not result in any substantial improvement in accuracy, and in fact tends to make the estimates even more negative. The similarity between the estimated and model DTR trends is not as good as would be expected from the high correlations listed in Table 3.2. In general, the estimated DTR trends made using all inputs are more negative than the actual model DTR trends, suggesting a systematic bias in the formulation of Eq. (3.4). This bias exists in the estimates made with only $S_c$. 

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Table 3.3: 1950-2100 trends in the DTR in the CGCM1 GHG+A1 integration estimated from Eq. (3.4). For each estimate, the time series from GHG+A1 is used for the indicated variable(s); other variables are assumed constant at the mean values in CGCM1 CTRL. Trends are in °C per century.

<table>
<thead>
<tr>
<th>Variable</th>
<th>66-33°S</th>
<th>33°S-0°</th>
<th>0°-33°N</th>
<th>33-66°N</th>
</tr>
</thead>
<tbody>
<tr>
<td>DTR</td>
<td>-0.9</td>
<td>0.4</td>
<td>-0.4</td>
<td>-1.1</td>
</tr>
<tr>
<td>$S_c$ (Solar radiation)</td>
<td>-1.7</td>
<td>-0.1</td>
<td>-0.7</td>
<td>-1.2</td>
</tr>
<tr>
<td>$\alpha_g$ (Ground albedo)</td>
<td>-0.5</td>
<td>-0.2</td>
<td>-0.5</td>
<td>-0.6</td>
</tr>
<tr>
<td>$H_0$ (Heat fluxes)</td>
<td>0.1</td>
<td>-0.4</td>
<td>0.5</td>
<td>0.2</td>
</tr>
<tr>
<td>$w_g$ (Soil moisture)</td>
<td>-0.3</td>
<td>-0.5</td>
<td>0.3</td>
<td>0.2</td>
</tr>
<tr>
<td>$T_0$ (Mean temperature)</td>
<td>-0.4</td>
<td>-0.2</td>
<td>-0.3</td>
<td>-0.5</td>
</tr>
<tr>
<td>$S_c, \alpha_g, H_0, w_g, T_0$</td>
<td>-2.6</td>
<td>-1.4</td>
<td>-0.7</td>
<td>-1.0</td>
</tr>
<tr>
<td>$S_c, w_g$</td>
<td>-2.0</td>
<td>-0.7</td>
<td>-0.4</td>
<td>-0.4</td>
</tr>
</tbody>
</table>

and $w_g$ as inputs, further suggesting that the bias involves one of these variables. It appears that either the effect of $S_c$ is being overestimated (since it tends to predict negative trends) or that the effect of $w_g$ is being underestimated (since it tends to predict positive trends).

We now turn to a statistical examination of the model output to resolve this question by showing that the effect of $w_g$ is in fact being underestimated.

### 3.5 Statistical Analysis

Calculations with Eq. (3.4) indicate that the radiative effects of clouds and the heat capacity of soil moisture are the controlling influences on DTR variability. However, the analytic model is quite simple and makes many approximations. The effects of this are most evident in the estimated DTR trends, which are generally more negative than the model changes. However, the analytic model was useful in providing a physical framework for this investigation. In this section we turn to linear regression models to confirm the importance of clouds and soil moisture to the DTR trends, and to resolve the question of the systematic bias found above. In the last section, clouds were found to be important for both the DTR variability and DTR trends. While soil moisture did not appear important for the trends, it was the only other variable highly correlated with DTR variability, so it follows that we
the following analysis in case the analytic equation was badly estimating the amplitude of
the DTR trends predicted using soil moisture. While these statistical models are not as
facilitating to physical insight as is the analytic model of Eq. (3.4), they may yield more
accurate results since their construction depends on the data themselves.

The regression models are constructed as the least squares estimate:

$$\Delta \hat{T}_G \sim \Delta T_C + X_G[(X_C^T X_C)^{-1}X_C^T(\Delta T_C - \Delta T_C)].$$

(3.5)

Here $\Delta \hat{T}_G$ is the estimated DTR time series, while $\Delta T_C$ is the DTR time series in the 201
years of the CGCM1 CTRL simulation. $\Delta T_C$ is the mean of $\Delta T_C$. $X_C$ is an $N_t \times N_k$
matrix containing the times series, of length $N_t$, of the $N_k$ input variables in the CGCM1
CTRL simulation. $X_G$ is the corresponding matrix from the global warming simulation.
The time series (columns) in both $X_C$ and $X_G$ are anomalies about the respective means in
the control simulation. The $T$ superscript denotes the transpose. Therefore, this statistical
model is constructed using the correlation between variables in CGCM1 CTRL, but is then
applied to the data from CGCM1 GHG+A1.

The correlation between the model DTR and the estimates from Eq. (3.5) using the
listed input variables are given in Table 3.4 for the DJF and JJA seasons in CGCM1
GHG+A1 over the four zonal bands. The test uses CGCM1 GHG+A1 output since the
model specifically finds the best fit to the CGMC1 CTRL output, and so this test is not
affected by possible overfitting. However, since rather large trends occur and dominate the
variance in many of the variables in CGCM1 GHG+A1, the second order least squares
polynomial fit is removed from all time series before the calculation of the correlation. As
required by the values in Table 3.2, estimates produced either by $S_c$ or by $w_g$ are highly
correlated with the DTR, and use of both variables generally improves the estimate. The
exceptions are of course the winter seasons at middle latitudes.

The DTR trends obtained in CGCM1 GHG+A1 according to these estimates are listed
Table 3.4: Correlations between the model DTR and linear regression estimates. Values are calculated from the output of CGCM1 GHG+A1 after removal of the second order least squares polynomial fit, and span four zonal regions. Seasonal means of the variables were averaged over the zonal bands before input into the equation. $S_{c_{res}} (w_{g_{res}})$ is the residual of $S_c$ ($w_g$) after removal of the component correlated with $w_g$ ($S_c$). See the text for a description of the regression estimation. Values outside of the -0.1 to 0.1 interval are significant at the 5% level assuming white noise processes.

<table>
<thead>
<tr>
<th>Variable</th>
<th>66-33°S</th>
<th>33°S-0°</th>
<th>0°-33°N</th>
<th>33-66°N</th>
</tr>
</thead>
<tbody>
<tr>
<td>$S_c$ (Solar radiation)</td>
<td>0.8</td>
<td>0.5</td>
<td>0.9</td>
<td>0.8</td>
</tr>
<tr>
<td>$f_{c_{res}}$ ($S_c$ residual)</td>
<td>0.3</td>
<td>0.3</td>
<td>0.6</td>
<td>0.5</td>
</tr>
<tr>
<td>$f_c$ (Cloud fraction)</td>
<td>0.7</td>
<td>0.3</td>
<td>0.9</td>
<td>0.8</td>
</tr>
<tr>
<td>$w_g$ (Soil moisture)</td>
<td>0.8</td>
<td>0.7</td>
<td>0.8</td>
<td>0.8</td>
</tr>
<tr>
<td>$w_{g_{res}}$ ($w_g$ residual)</td>
<td>0.6</td>
<td>0.7</td>
<td>0.4</td>
<td>0.5</td>
</tr>
<tr>
<td>$S_{c_{res}}, w_{g_{res}}$</td>
<td>0.9</td>
<td>0.7</td>
<td>1.0</td>
<td>0.9</td>
</tr>
</tbody>
</table>

In Table 3.5. As expected, use of $S_c$ produces similar trend estimates to those in Table 3.3 using the analytic model with $S_c$. However, the trends estimated with $w_g$ are generally much larger than the corresponding values in Table 3.3, suggesting that the bias in Eq. (3.4) lies in an underestimation of the effect of soil moisture. Indeed, the combined use of the two predictands $S_c$ and $w_g$ produces estimates very similar to the model trends, with the not surprising exception of the winter seasons trends over the middle latitudes. The global annual mean 1950-2100 variations in the DTR estimate based on solar radiation and soil moisture are shown in Figure 3.5. The strong similarity to the model DTR variations is consistent throughout the time period. The corresponding trends at each grid point are displayed in Figure 3.6.

From Eq. (3.3), $S_c$ is a function of both $f_c$ and $\alpha_c$. Since the latter is not archived from the CGCM1 integrations we examine $f_c$ to determine the cause of the changes in $S_c$. In fact, $S_c$ and $f_c$ are very highly correlated and are comparable in their relation to the DTR (Table 3.4). However, the estimated trends in the DTR in CGMC1 GHG+A1 differ systematically by about 1°C/century (Table 3.5). The increased absorption of solar near infrared radiation by more water vapour in a warmer atmosphere has been suggested as a
Table 3.5: Trends in the DTR and estimates from linear regression models. Values are in °C/century, span four zonal regions, and are calculated from the output of CGCM1 GHG+A1. $S_{res}$ and $w_{gres}$ are defined as in Table 3.4. See the text for a description of the regression models.

<table>
<thead>
<tr>
<th>Variable</th>
<th>66-33°S</th>
<th>33°S-0°</th>
<th>0°-33°N</th>
<th>33-66°N</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>DJF</td>
<td>JJA</td>
<td>DJF</td>
<td>JJA</td>
</tr>
<tr>
<td>DTR</td>
<td>-0.9</td>
<td>0.4</td>
<td>0.3</td>
<td>0.2</td>
</tr>
<tr>
<td>$S_c$ (Solar radiation)</td>
<td>-1.0</td>
<td>0.0</td>
<td>-0.5</td>
<td>-0.6</td>
</tr>
<tr>
<td>$f_c$ (Residual)</td>
<td>-0.4</td>
<td>0.0</td>
<td>-0.7</td>
<td>-0.5</td>
</tr>
<tr>
<td>$w_g$ (Soil moisture)</td>
<td>-0.5</td>
<td>0.0</td>
<td>0.9</td>
<td>0.6</td>
</tr>
<tr>
<td>$w_{gres}$ (水分残留)</td>
<td>0.0</td>
<td>-0.6</td>
<td>0.7</td>
<td>0.7</td>
</tr>
<tr>
<td>$S_c$, $w_g$</td>
<td>-0.9</td>
<td>-0.7</td>
<td>0.2</td>
<td>0.1</td>
</tr>
</tbody>
</table>

Figure 3.5: Regression model estimate of annual mean variations in the DTR from CGCM1 GHG+A1 averaged over non-polar land areas. The regression time series, the black solid line, is estimated using downward solar radiation and soil moisture. The model DTR variations, the grey dashed line, are shown for comparison.
Figure 3.6: Map of the 1950-2100 annual mean trends in the DTR in CGCM1 GHG+A1 estimated from the regression model using downward solar radiation and soil moisture. A map of the actual model DTR trends is shown for comparison.
partial forcing of the DTR trend (Stenchikov and Robock 1995), and would manifest itself here as a trend in $S_0$. However, the expected magnitude of this effect is too small to account for the discrepancy here. That leaves a trend in $\alpha_c$ or in the diurnal cycle of cloud cover as possible causes. The increase of precipitation over land in CGCM1 (Boer et al. 2000b) indicates optically thicker clouds, which suggests that an increase in $\alpha_c$ is plausible.

Soil moisture and solar radiation are strongly related at the seasonal time scales examined here. Soil moisture is determined by precipitation and evaporation, both of which are related to clouds and thus solar radiation. Conversely, clouds depend upon the availability of water vapour, which in turn depends on the soil moisture. Thus the question arises as to the relative importance of both variables. To examine this we decompose $S_c$ and $w_g$ into

$$S_c = S_{corr} + S_{res}$$

$$w_g = w_{g,corr} + w_{g, res}$$

where $S_{res}$ ($w_{g, res}$) is the residual after the removal of the component $S_{corr}$ ($w_{g, corr}$) correlated with $w_g$ ($S_c$).

The correlations of the DTR estimates of these residuals with the model DTR in CGCM1 GHG+A1 are listed in Table 3.4. The two residuals account for comparable portions of the variance of the DTR, with $S_{res}$ being slightly more important. The magnitudes of the DTR trend estimates based upon each of the residuals (Table 3.5) are also quite similar. The equal importance of solar radiation and soil moisture for the DTR trend found here differs from the predominance of solar radiation suggested by the analytic model. This could reflect the separation in Eq. (3.4) of soil moisture from the ground albedo and latent heat flux, both of which depend strongly upon it. However, in the model integrations changes in the latent heat flux are generally balanced by the sensible heat flux. Inherent in this point, of course, is the assumption that both heat fluxes exhibit similar diurnal cycles. Furthermore, when variations in the ground albedo and latent heat flux are included in Eq. (3.4), the estimates in fact generally worsen. A plausible explanation is the substitution of the screen
temperature for the surface temperature in Eq. (3.4), which systematically underestimates the diurnal amplitude of $T$ (Peel 1974, Jin et al. 1997) and thus the magnitude of its effect on the DTR.

### 3.6 Middle Latitude Winter

The results of the previous two sections indicate that in most seasons and locations the DTR trends result primarily from combined changes in clouds and soil moisture. However, this relation does not hold for the middle latitude winters. This discrepancy is important, since the largest DTR decrease occurs at this time. We examine this case more closely in this section.

The daily average DTR as a function of the daily mean temperature is displayed in Figure 3.7 for the DJF season in the northern middle latitudes. Daily values at each grid point over the 1961-1980 period from CGCM1 GHG+A1 are used. The most striking feature of this figure is the much lower DTR values that occur on days when the mean temperature is near 0°C. As noted by Kharin and Zwiers (2000), this is an artifact of CGCM1’s land surface model. (CGCM2 also uses this land surface scheme and thus is similarly impacted.) CGCM1 represents the land surface as a single layer. Due to energy balance constraints, the temperature near the surface does not drop substantially below the freezing point until all of the soil moisture in the layer freezes. Thus, on days with a mean temperature near 0°C, the energy flux into the ground is often insufficient to completely freeze or melt soil moisture, resulting in little change in the screen level temperature over the diurnal cycle.

During the 1961-1980 period, most winter days and grid boxes in the northern middle latitudes are below the freezing point, with the average being -11.8°C (Figure 3.8). However, these latitudes warm by 7.9°C by the 2081-2100 period, resulting in a much higher frequency of days and grid boxes near 0°C. This implies a higher frequency of days with a very small DTR. We can estimate the effect on the DTR of this shift to more frequent days near the
Figure 3.7: The daily average DTR as a function of the daily mean temperature. 1961-1980 daily data for the DJF season, covering grid boxes in the northern middle latitudes, are used from CGCM1 GHG+A1.

freezing point by integrating the product of the mean temperature \( T_0 \) probability density function of 2081-2100 in Figure 3.8 with the DTR-\( T_0 \) function from 1961-1981 in Figure 3.7. The result is a 1.3°C decrease in the DTR. The actual model DTR decrease between these two time periods is 1.4°C, so this shift to more frequent days and grid boxes near the freezing point accounts for most of the change in the DTR.

During the southern middle latitude winter, this shift in mean temperature near 0°C is also important. However, since most of the land here is near the low latitudes and the mean temperature is usually above freezing, this leads to an increase in the DTR which is moderated by the influence of clouds and soil moisture. Since the mean temperature seldom reaches the freezing point in the other seasons in either the northern or southern middle latitudes, the DTR is not as susceptible to this effect, and so the trends arise primarily as a consequence of changes in clouds and soil moisture. The same applies to all seasons in the low latitudes.
Figure 3.8: The frequency of days and grid boxes with given mean temperatures during the DJF season in the northern middle latitudes. The frequencies for the 1961-1980 and 2081-2100 periods in CGCM1 GHG+A1 are shown.

3.7 Discussion and Conclusions

Recent modelling studies support the theory that anthropogenic forcing of climate will induce a decrease in the DTR. However, the predicted magnitude of this trend is much smaller than observed. Here we have examined the trends in global warming integrations of a coupled GCM, and found decreases similar to those in previous studies. Trends tend to be largest in the middle latitudes, and possess distinct seasonal signatures, although these are different in either hemisphere.

Variations in the DTR are considerably more sensitive to forced changes in other aspects of the climate system rather than directly to the forcings themselves. In particular, projected decreases result from the combined effects of changes in clouds and in soil moisture, consistent with the findings of previous observational studies (Karl et al. 1993, Dai et al. 1997a, 1999) and modelling studies (Stenchikov and Robock 1995, Dai et al. 2001). The main impact of clouds is to reduce the downward solar radiation during the day, and thus reduce $T_{max}$. Thus the reduction in solar radiation in the CGCM1 GHG+A1 integration arises either through an increase in the mean cloud albedo or through a shift in the diurnal cycle of cloud cover, since the average cloud cover changes little. The former alternative is
supported by the increase in precipitation occurring in the integration, since this implies thicker and more reflective clouds, and it is supported by observational studies (Dai et al. 1997a). Notably, the scattering of sunlight by sulphate aerosols is found to have a negligible impact on the DTR, as found in other modelling studies (Mitchell et al. 1995, Reader and Boer 1998). It should be noted however that this coupled model lacks a representation of the indirect effect of aerosols; since this effect modifies the optical properties of clouds, it could have an important influence on the DTR trends.

The exceptions to the relation of the DTR to solar radiation and soil moisture occur during the winter seasons in the middle latitudes. The reflective and insulative properties of snow reduce the effect these variables would otherwise have. Nevertheless, the largest decrease in the DTR occurs at this time. There is a tendency in CGCM1 for the DTR to be very small when the mean temperature is near the freezing point. The DTR trends during the winter result from a shift to a higher or lower frequency of days near the freezing point, and thus with very small DTR values. It should be noted that this effect is an artifact of the land surface scheme in CGCM1 (and CGCM2) and so the affected trends cannot be considered realistic.

Analysis of the effects of soil moisture on the DTR trends indicate that its influence is comparable to that of clouds in the model integrations. Much of this effect arises through its control of the ground heat capacity, although the comparison of the analytic and regression models in Sections 3.4 and 3.5 indicates that this can probably not fully account for the response. Other influences arise through control of the ground albedo and the latent heat flux. While both of these were represented in the analytical model, the representation of the latent heat flux assumed a similar diurnal cycle to the sensible heat flux, and the anti-correlation between the two heat fluxes resulted in a cancellation of their effects, consistent with the findings of observational studies (Dai et al. 1999). However, both variables are individually highly correlated with the DTR, and so relatively small deviations from this assumption may be important enough to give rise to the discrepancy in the effect of soil
moisture. While the use of a single layer model for the land surface scheme limits the applicability of using the coupled model output for predictive purposes, it does indicate the potential for an important influence of land surface processes on the DTR. For example, this implies that the physiological responses of vegetation to climate change, which are not represented in CGCM1, could be important in determining the tendency of the DTR, as proposed by Collatz et al. (2000). Also, although observational studies suggest that changes in land use are not a primary factor behind the observed DTR trend (Easterling et al. 1997, Gallo et al. 1999), the importance of soil moisture found here implies that this effect requires further observational and modelling study.

These results indicate that anthropogenic forcing could induce a decrease in the DTR, through changes in clouds and in land processes. Naturally, these results depend on the ability of the coupled model to represent these two components, whose behaviour under climate change is amongst the least understood of all processes. Consequently, much of the current activity in model development is concentrated on these two factors. The resulting improvements should soon permit a more confident determination of the behaviour of the DTR under anthropogenic forcing. At this stage, however, it is possible to identify these two factors as the primary influence on the DTR under climate change.
Chapter 4

Projection of Climate Change onto Modes of Atmospheric Variability

4.1 Introduction

A basic issue in climate studies is the nature of the response of the climate system to the enhanced radiative forcing produced by increasing atmospheric concentrations of greenhouse gases. Recently there has been considerable interest in the possibility that this change may project onto the pre-existing natural modes of variability of the climate system. However, this projection could take one of several different forms, depending, for instance, upon whether the dynamics of the change are linear or nonlinear. The climate change could also project onto several modes or alternatively just a single pattern of variability. Knowledge of the existence and nature of such projections would greatly enhance our ability to both detect and project climate change. In an attempt to expand this knowledge, this chapter consists of a simple evaluation of two interpretations of this projection based on linear and nonlinear perspectives of the climate system, using integrations of a coupled general circulation model (GCM).

Support for hypotheses of the projection of enhanced greenhouse warming onto the dominant modes comes from analyses of both the observational record and GCM output. For example, evidence of recent trends in the Northern Annular Mode (NAM, also known as the Arctic Oscillation) (Thompson and Wallace 1998, 2000) and the Southern Annular
Figure 4.1: Comparison of linear and nonlinear interpretations of the projection of climate change. Suppose a hypothetical mode of climate variability whose amplitude has the multi-modal PDF shown in a). Climate change could project b) linearly onto this mode through a translation of the PDF, or c) non-linearly through a change in the shape of the PDF, in this case a shift in the residence frequency of the two regimes associated with this mode. Note that both types of projection result in shifts in the mean of the PDF, indicated by the chevron.

Mode (SAM, also known as the Antarctic Oscillation) (Rogers and van Loon 1982, Nigam 1990, Thompson and Wallace 2000) suggest that they represent substantial fractions of the change in the respective hemispheric climate (Thompson and Wallace 1998, Thompson et al. 2000, Thompson and Solomon 2002). In transient integrations of their respective GCMs Fyfe et al. (1999) and Shindell et al. (1999) both noted trends in the mode of sea level pressure (SLP) representing the NAM, corresponding in sign to the observed trend. Fyfe et al. (1999) and Kushner et al. (2001) also observe similar trends in the simulated SAM, and find that these trends in fact represent almost the entire Southern Hemisphere SLP response.

In interpreting these projections of climate change onto the dominant modes of variability, the simplest case views the climate system from a linear perspective. In this interpretation, the climate change manifests itself through a translation of the mean state of the mode (Figure 4.1). For example, a linear translation in the SAM could result from a linear change in the Antarctic Ocean-Antarctica temperature gradient. Support for this interpretation comes from the analyses of monthly GCM output of Fyfe et al. (1999) and Shindell et al. (1999), which suggest translations of Gaussian variables.

On the other hand, a nonlinear perspective of the climate system can lead to a different
interpretation of the projection. Palmer (1999) considers the case of a weak forcing on a simple nonlinear system as an analogy to enhanced greenhouse forcing on the climate system, and suggests that the resulting climate change would be reflected in a shift of the residence frequency of the system in certain quasi-stationary regimes. This shift would be visible as a change in the shape of the probability density function (PDF) of the principal components (PCs) associated with these regimes (Figure 4.1). However, as a consequence of the stability of the attractors, the location and physical structure of these regimes would remain constant, and thus the structure of the corresponding empirical orthogonal functions (EOFs, the spatial patterns corresponding to the PCs) would remain relatively unaffected.

Corti et al. (1999) examine this interpretation in the Northern Hemisphere mean monthly extended winter 500 hPa geopotential height from reanalysis output, covering the latter half of the twentieth century. They detect multi-modality in the system, although Hsu and Zwiers (2001) find that only one of the regimes, corresponding to the so-called "cold ocean and warm land" (COWL) pattern (Wallace et al. 1996), is robust. Noting changes in the structure of this multi-modality over time, Corti et al. (1999) suggest that climate change experienced over this period was reflected in a shift in the residence frequency of the climate system in these regimes.

Hsu and Zwiers (2001) also analyse the 500 hPa geopotential height field, but from equilibrium integrations of a coupled GCM, and find that while the locations of the robust regimes are insensitive to the greenhouse forcing, the occupation statistics change, consistent with the nonlinear interpretation. Monahan et al. (2000a) also examine extratropical Northern Hemisphere sea level pressure (SLP) from these same equilibrium integrations using a nonlinear PC analysis, and find substantial changes in the occupation statistics of the leading nonlinear PC in a warmer climate. Furthermore, Monahan et al. (2000b) find that the recent trend in the observed NAM is better defined as a change in the occupation statistics of associated quasi-stationary regimes.

In this nonlinear perspective of the atmospheric system, fluctuations between different
regions of the atmospheric attractor occur on time scales of around one week. However, such a perspective can be altered to reflect time scales representative of the coupled ocean-atmosphere climate system. Evidence for regime-like behaviour on these time scales exists in the form, for example, of the Pacific Decadal Oscillation (PDO) (Mantua et al. 1997). Of course, some modes, such as the NAM, are predominantly atmospheric in nature and thus likely do not exhibit regime behaviour on these interannual time scales.

This chapter aims to evaluate both of these linear and nonlinear interpretations of enhanced greenhouse warming projecting onto the dominant modes of variability, on the longer time scales representative of coupled ocean-atmosphere dynamics. We do this by examining the differences between annual mean SLP and surface air temperature (SAT) fields from 1000 year equilibrium integrations of the Geophysical Fluid Dynamics Laboratory (GFDL) coupled GCM run with different greenhouse gas forcing. These surface fields are used since it is at this level that the main components of the climate system interact. Of course, real changes in radiative forcing occur gradually, and so conclusions derived from these equilibrium fields are further tested on the same fields from two transient integrations of the same GCM.

The outline of the rest of this chapter is as follows. Section 4.2 consists of a brief description of the GFDL coupled model and the output fields, while Section 4.3 contains a brief outline of the methods used to analyse the variability of these fields. The nature of the dominant modes of variability are examined in Section 4.4. Since the linear interpretation being considered depends upon the physical structure of these modes remaining invariant in warmer climates, the stability of the modes between different climates is also tested. In Section 4.5 we study the spatial projection of the mean changes in the output fields under different greenhouse gas forcing onto the dominant modes of variability, thus testing whether the climate change projects onto the dominant modes. We then examine the PDFs of the climate system in the space spanned by the corresponding PCs in Section 4.6 in order to differentiate between the linear and nonlinear interpretations. Finally, Section 4.7 consists
of a discussion of the results from the previous sections.

4.2 The GFDL Model Output

We give only a brief description of the GFDL coupled model; for more details see Manabe et al. (1991) and Stouffer and Manabe (1999). It consists of general circulation models of the atmosphere and ocean, as well as land surface and sea ice models. Its domain is global and its geography is realistic within the limits of the resolution.

The atmospheric component has nine vertical finite difference levels with horizontal distributions of atmospheric variables represented by spherical harmonics (15 Legendre functions associated with 15 Fourier components) and by grid-point values (Gordon and Stern 1982). Insolation varies seasonally, but not diurnally. Cloud cover is predicted based on relative humidity, while a simple land surface model is used to compute surface fluxes of heat and water (Manabe 1969).

The oceanic component (Bryan and Lewis 1979) employs a full finite difference technique and uses a regular grid system with approximately 4.5° latitude by 3.7° longitude spacing. There are 12 vertical levels in the ocean. A simple sea ice model is also included which computes sea ice thickness based on thermodynamic heat balance and advection of sea ice by ocean currents. The atmospheric, oceanic and sea ice components exchange fluxes of heat, water and momentum once per day.

The coupled model’s quasi-equilibrium initial condition is obtained by separate integrations of the atmospheric and oceanic components using the observed surface boundary conditions. When the integration begins from this initial state, the model drifts towards its own less realistic state. To prevent this drift, the fluxes of heat and water are modified by flux adjustments. These adjustments are determined prior to the coupled integration and do not change through the course of the integration; thus they neither systematically damp nor amplify the surface anomalies that occur during the course of the integration.
Although the adjustments do not eliminate the shortcomings of the model (Marotzke and Stone 1995), they do help prevent the rapid drift of the model from its initial state.

Three multiple-millennia integrations have been performed using this coupled model. In the first, the control integration, no changes were made to the radiative forcing in the model. In the other two integrations the CO$_2$ concentration in the model's atmosphere increased at a rate of 1% per year to doubling (2×CO$_2$) and quadrupling (4×CO$_2$)). These two increased-CO$_2$ integrations were then run to equilibrium with the radiative forcing held constant at 2× and 4×CO$_2$, respectively. In this model, it takes about 4000 years to reach a statistical equilibrium.

For the analysis shown here, the annual mean SAT and SLP fields from 1000 year segments are used from all three integrations. The use of 1000 year time series allows very accurate sampling of variability on time scales shorter than 100 years. Output from two transient integrations of this model, with CO$_2$ levels rising at 1% per annum from 1× to 2× (70 years) and from 1× to 4× present levels (140 years), are also examined to confirm the results in changing climates.

4.3 Method

The climate modes examined in this analysis are defined as the leading EOFs of the output fields, using extratropical hemispheric (poleward of 20° latitude) or global domains where appropriate. EOFs in the control integration are retained provided they are clearly separated according to the criterion of North et al. (1982). If “real” dynamical modes are important contributors to global variability, higher order EOFs will represent the same variability, albeit perhaps via linear combinations rather than individual patterns. Furthermore, there is no a priori reason to suppose that climate change will project onto the leading mode and not the lesser ones. Therefore, higher order EOFs are retained, with the understanding that projection of climate change onto the dominant modes may be de-
ected without the possibility of identifying the exact mode. The analysis is thus limited in the ability to identify actual physical modes. Furthermore, since EOFs may represent collections of physical modes, slight changes in the relative dominance of those modes could lead to rather different EOFs in warmer climates even though little change has occurred in the behaviour of the modes themselves. Nevertheless, PC analysis represents a viable quantitative technique for this investigation.

In standard PC analysis areas with larger variance, for instance SAT at the sea ice edge, tend to dominate the EOFs. This drowns out patterns of variability in areas with smaller variance which may nevertheless be important for global climate. A solution is to standardise the data fields (i.e. divide by the local estimates of the standard deviation) before conducting the PC analysis. In this study we examine the EOFs of both the output fields and their standardised versions, hereafter referred to as the physical output fields and the standardised output fields, respectively. The physical EOFs tend to focus on higher latitudes, since the variance in SLP and SAT is largest there; the lower latitude variability is relatively more important in the standardised EOFs.

To examine the possible importance of regime behaviour, we examine the SLP and SAT fields in a reduced space in the same manner as Kimoto and Ghil (1993), Brunet (1994), Corti et al. (1999), Hsu and Zwiers (2001) (see Section 4.6). The two-dimensional PDFs in the space spanned by the leading two PCs are estimated using an adaptive kernel estimator (Silverman 1986, Kimoto and Ghil 1993). This non-parametric method uses all of the values in the PCs, weighted according to a kernel function (here the Gaussian kernel), to estimate the value of the PDF at each point. The optimal smoothing parameter, \( h \), of this estimator, which determines the effective radius of influence of the kernel function, is determined as the location of the minimum of a score function \( m(h) \).
4.4 The Dominant Modes of Variability

In order for the linear interpretation of climate change projecting onto the dominant modes of variability to be viable, these modes must remain important in the warmer climates. However, this requirement may not always hold in the nonlinear interpretation evaluated here, since if the climate system tends towards residing overwhelmingly in only one regime, fluctuations between this regime and others will occur less frequently and thus the mode of variability associated with these fluctuations will be less important. This section presents the dominant modes of variability of the SLP and SAT fields extracted using PC analysis, and examines whether they remain important in climates with higher greenhouse gas concentrations.

The EOFs of physical SLP are presented first in the domain of the extratropical hemispheres; those in the global domain closely resemble their hemispheric counterparts and so are ignored in this analysis. The subsequent subsection examines the EOFs of standardised SLP over the global domain, which brings out patterns in the Tropics, unlike the EOFs of physical SLP. Since the Tropics dominate the EOFs of standardised SLP, we do not look at the hemispheric domains in this case as these do not contain all of the Tropics. This ordering is then repeated for the EOFs of SAT.

4.4.1 Extratropical Northern Hemisphere Physical SLP

The four leading EOFs of extratropical Northern Hemisphere SLP (Figure 4.2) all satisfy the criterion of North et al. (1982). The first EOF, resembling the NAM (Thompson and Wallace 1998, 2000), and EOFs 2 and 3 are quite similar to the three leading EOFs of extended winter monthly mean SLP found by Fyfe et al. (1999) in a control integration of the Canadian Centre for Climate Modelling and Analysis (CCCma) GCM and in the observational record. The main differences in our EOFs are the strong variability in the North Pacific, as noted by Hall and Visbeck (2002), and the importance of the Eurasian
centre in EOFs 2 and 3. In the control integration these patterns represent 23%, 9%, 8% and 6% of the variance, respectively. The first EOF remains practically unchanged in the warmer climates, while the second and third EOFs appear to mix in the warmer climates but nevertheless remain important (Figure 4.3). The intensities of the centres in the fourth EOF vary in the warmer climates, but the basic tripole pattern remains.

4.4.2 Extratropical Southern Hemisphere Physical SLP

The first EOF of extratropical Southern Hemisphere SLP (Figure 4.4) resembles the SAM (Rogers and van Loon 1982, Nigam 1990, Thompson and Wallace 2000), while the second EOF (Figure 4.2) resembles an average of the second EOFs of SLP found in the observational record by Rogers and van Loon (1982) in the winter and summer seasons. The top three EOFs represent 41%, 7% and 5% of the variance in the control integration, respectively. The first EOF remains practically unchanged in the warmer climates, while the centre in the second EOF weakens and shifts westward in the warmer climates, but continues to remain important. The third EOF, however, looks rather different in the warmer climate fields, although the pattern still accounts for an important portion of the variability in these climates.

4.4.3 Global Standardised SLP

The first six EOFs (Figure 4.5) of global standardised SLP all satisfy the criteria outlined in Section 4.4.1. The first and fourth EOFs are similar to the SAM and NAM, respectively, but the middle latitude band extends over the Tropics in both cases, reflecting the stronger role of this region in this standardised domain. The importance of the Tropics is further illustrated in the second EOF, which resembles the Southern Oscillation (SO). Of particular note, the first EOF appears to partially represent a Tropics-to-poles mode of variability along with the SAM. This dual representation implies that neither mode is fully represented, resulting in "echoes" of this structure in EOFs 3, 4, and 5. These six patterns represent
Figure 4.2: The four leading EOFs of extratropical Northern Hemisphere physical SLP in the control integration.
Figure 4.3: The four leading EOFs of extratropical Northern Hemisphere physical SLP in each of the equilibrium simulations ($1 \times \text{CO}_2$, $2 \times \text{CO}_2$, and $4 \times \text{CO}_2$). The fraction of the variance represented by these patterns is given at the top of each plot. The colour scale is the same as in Figure 4.2.
The three leading EOFs of extratropical Southern Hemisphere physical SLP in the control integration.

12%, 9%, 6%, 5%, 4%, and 3% of the standardised variance, respectively. All of the patterns retain their overall structure in the warmer climates, with the largest differences related to the magnitude of the Tropics-to-poles structure (not shown).

**4.4.4 Extratropical Northern Hemisphere Physical SAT**

Only the first EOF of extratropical Northern Hemisphere physical SAT (Figure 4.6), representing 10% of the variance in the control integration, clearly satisfies the criterion of North et al. (1982). Due to the relatively small variance of SAT in the Tropics, this EOF closely resembles the top EOF of SAT north of 50°S found by Barnett (1999) in a separate 1000 year segment of the same multiple-millennia integration used here. The Eurasian centre shifts eastwards in the warmer climates, but the main tripole pattern is maintained (not shown).
Figure 4.5: The six leading EOFs of global standardised SLP in the control integration.

Figure 4.6: The leading EOF of extratropical Northern Hemisphere physical SAT in the control integration.
4.4.5 Extratropical Southern Hemisphere Physical SAT

The largest variations in physical SAT in the extratropical Southern Hemisphere occur around Antarctica, and consequently the leading EOFs are concentrated in this region (Figure 4.7). The PC time series associated with the first EOF is shown in Figure 4.8. This pattern represents oscillations between two states with temperatures west of the Antarctic Peninsula 6°C apart. Considering its location, this pattern probably represents fluctuations between ice-covered and ice-free regimes with associated changes in ocean convection. The second PC time series similarly oscillates between two regimes (not shown), but the spatial pattern of the corresponding EOF suggests that it represents shifts in the location of sea ice and ocean convection. Consequently, we describe both of the patterns as "sea ice modes." The three leading EOFs represent 27%, 14%, and 6% of the variance in the control integration. Since these patterns are related to changes in sea ice extent, they do not remain important in the warmer climates (not shown), when the ice has retreated. Thus, only the nonlinear interpretation of climate change projection can be examined in this domain in later sections.

4.4.6 Global Standardised SAT

The three leading EOFs of global standardised SAT (Figure 4.9) all satisfy the separation criterion of North et al. (1982). The first EOF represents an El Niño/La Niña-like warming and cooling in the Tropical Pacific Ocean. At least partly due to the coarse ocean resolution, the El Niño/Southern Oscillation (ENSO) simulated by the GFDL model is weaker than the observed (Knutson and Manabe 1994), and so this pattern is not detected in physical SAT. The second EOF resembles the first EOF of Southern Hemisphere physical SAT, while the third pattern bears a close resemblance to the top EOF of extratropical Northern Hemisphere physical SAT. These three patterns represent 7%, 4% and 3% of the variance of global standardised SAT in the control integration respectively. Since it represents sea
Figure 4.7: The three leading EOFs of extratropical Southern Hemisphere physical SAT in the control integration.

Figure 4.8: The PC time series associated with the leading EOF of extratropical Southern Hemisphere physical SAT in the control integration.
ice variability, the second EOF weakens as the sea ice retreats, and so only the first and third EOFs remain important in the warmer climates (not shown). Knutson and Manabe (1994) also note that ENSO variability, represented here by the first EOF, does not change in the warmer climates.

4.4.7 Summary

The leading EOFs of SLP collectively account for an important fraction (40-50\%) of the variability in all domains. They also generally remain important in the warmer climates, with only relatively small changes occurring in their structure. On the other hand, many of the top EOFs of SAT represent variations in sea ice extent; with the retreat of the sea ice edge these modes cease to remain important in the warmer climates. However, the top SAT EOFs unrelated to sea ice variability do remain important in the warmer climates. Nevertheless, they account for a smaller portion (~10\%) of the variability. Since they remain important in the warmer climates, SAT patterns unrelated to sea ice variability, as well as SLP patterns, may be consistent with the linear or nonlinear interpretations of climate change projecting onto the dominant modes of variability. On the other hand, those SAT patterns associated with sea ice variability, which changes considerably in the warmer
climates, may be consistent with only the nonlinear interpretation.

4.5 **Spatial Projection of Climate Change**

Since EOFs represent the preferred direction of variability of the climate system, forced climate change may project onto these modes. In this section we test this possibility by examining the spatial projection of the climate change onto these patterns. We do this by taking the mean difference between the $2\times$ and $4\times$CO$_2$ equilibrium integrations and the control integration and then calculating the spatial correlation of these difference fields with the leading EOFs of the control integration. Of course, the actual rise in greenhouse gas concentrations is occurring gradually; thus the patterns of the mean trends in the output of the two transient integrations are also projected onto the control EOFs to confirm or differentiate the results in changing climates. A point to note is that both the difference fields and the EOFs often have non-zero means, since they are either defined in hemispheric domains or they are not constrained globally. This implies that the projection reflects both similarities in the pattern and in the direction of the change.

4.5.1 **Extratropical Northern Hemisphere Physical SLP**

The mean changes in physical SLP under enhanced greenhouse forcing are displayed in Figure 4.10. A notable feature of these changes in the Northern Hemisphere is their structural similarity, reflecting a fairly linear change in the global distribution of mean SLP in the warmer climates. The projection of these difference fields onto the top EOFs of physical SLP are plotted in Figure 4.11. It must be reminded that these projections reflect both the pattern and direction of the change. Roughly one third of the change projects onto the NAM-like mode. Fyfe *et al.* (1999) note a similar projection onto the top EOF of wintertime monthly mean SLP in the Northern Hemisphere in a transient integration of the CCCma coupled model, although they also find a comparable projection onto their third
**Figure 4.10:** The change in physical SLP occurring under enhanced greenhouse forcing. The mean difference between the a) 2×CO₂ and b) 4×CO₂ equilibrium integrations and the 1×CO₂ control integration are shown, along with the total integrated trend in the c) 1-2×CO₂ and d) 1-4×CO₂ transient simulations.

EOF. Instead, we see a similar projection onto the second EOF, which is actually related more to the general decrease in pressure in the Northern Hemisphere than to the pattern of the change.

**4.5.2 Extratropical Southern Hemisphere Physical SLP**

In the Southern Hemisphere the mean change in physical SLP between the equilibrium integrations projects (Figure 4.10) only partially onto the top modes (Figure 4.11). However, the mean trends in the two transient integrations differ substantially, projecting quite strongly onto the SAM-like mode. This difference may result from the delay of the sea
Figure 4.11: The spatial projection (squared spatial correlation) of enhanced greenhouse-forced climate change onto the top EOFs of the control integration. Both the difference fields and EOFs often have non-zero means, so the projection reflects both the pattern and direction of the change. The “p” and “s” prefixes denote “physical” and “standardised” versions of the data (SLP and SAT), respectively. The EOFs in each domain are labelled by their rank; see Figures 4.2-4.9 for depictions of each EOF.

Ice retreat and ocean warming in the Antarctic Ocean in the transient integrations. These results bear a striking resemblance to those of Fyfe et al. (1999) who find that in their transient integration of a coupled GCM the change in the Southern Hemisphere projected almost entirely onto the SAM-like pattern. Kushner et al. (2001) also observe this projection onto the SAM in the output of a higher resolution version of the GFDL coupled GCM than that used here. The implication of the difference between the equilibrium and transient changes here is that this is purely the initial response to climate change, with the system reverting to its original state once the climate system has evolved further.

4.5.3 Global Standardised SLP

As with physical SLP, the changes in standardised SLP (Figure 4.12) are structurally similar for the 2x and 4xCO₂ climates. However, the differences are not confined to the high
latitudes, instead reflecting a land-ocean difference. This change projects partly onto the third EOF (tropical Pacific-North Atlantic dipole), and negligibly onto the other modes (Figure 4.11). As with physical SLP, the changes in the transient integrations differ from that between the equilibrium integrations in the Southern Hemisphere, with the trend projecting fairly strongly onto the SAM-like mode, represented by the first EOF.

4.5.4 Extratropical Northern Hemisphere Physical SAT

As with SLP, the spatial structure of the change in physical SAT in the Northern Hemisphere (Figure 4.13) is similar in all integrations. It is also concentrated in high latitudes, due to the large ice albedo feedback. Nevertheless, this pattern projects negligibly onto the top
Figure 4.13: As in Figure 4.10 but for physical SAT.

EOF of this domain (Figure 4.11).

4.5.5 Extratropical Southern Hemisphere Physical SAT

In the Southern Hemisphere about 40% of the change in physical SAT between the equilibrium integrations (Figure 4.13) projects onto the first EOF (Figure 4.11). This is related both to the pattern and direction of the change, although it does not account for the largest warming in the $1 \times$ to $4 \times$CO$_2$ transient simulation, which occurs in the Ross Sea. Due to the lag in the ocean warming and sea ice retreat in the transient integrations the mean trends are much smaller than the equilibrium changes.
4.5.6 Global Standardised SAT

The changes in standardised SAT (Figure 4.14) are structurally quite similar for the 2x and 4xCO₂ climates, with the largest differences over the Tropics, but they differ from the trends in the transient integrations in high southern latitudes. In all integrations the change projects fairly strongly onto the ENSO-like mode (Figure 4.11); however, this is strongly reflective of the general warming. Furthermore, the pattern of change itself reflects simply a difference between the tropical and high latitudes and does not resemble the ENSO pattern at smaller scales, and thus a projection onto the ENSO-like mode would appear to be largely spurious.
4.5.7 Summary

The analysis of Northern Hemisphere physical SLP supports the results of Fyfe et al. (1999) that climate change will project only weakly onto the dominant modes in the Northern Hemisphere. Furthermore, the strong projection of climate change onto the SAM-like mode in the Southern Hemisphere found by Fyfe et al. (1999) and Kushner et al. (2001) is reproduced here in the transient integrations. However, the negligible projection between equilibrium integrations suggests that this constitutes only the initial response, and disappears once the sea ice has retreated and the ocean has warmed. The projection of the enhanced greenhouse-forced change in SAT onto the SAT modes is generally more representative of the overall warming rather than the spatial pattern of that warming. Inspection of Figure 4.11 shows that the change tends to project most strongly onto the more dominant modes. While this may be related in part to the greater spatial coherence of the more important EOFs, it nevertheless suggests that climate change will indeed tend to project onto the more dominant modes. Only in one case does this actually constitute most of the response, but it would be naive to assume that the simple implementation of linear PC analysis of annual mean output used here has extracted all of the important modes in the climate system.

4.6 The Nature of the Spatial Projection

The results of Section 4.5 demonstrate that forced climate change can project onto some of the dominant modes of variability. However, they do not differentiate between whether that projection is linear or nonlinear, that is whether the projection reflects a linear translation in the climate system or a shift in the residence frequency of the system in regimes associated with these modes. This section aims to distinguish between these two possibilities by examining the PDFs of the climate system in reduced spaces spanned by the top PCs of
the control integration.

4.6.1 Extratropical Northern Hemisphere Physical SLP

The estimated PDFs of physical SLP in the space spanned by the top two PCs of the control integration in the extratropical Northern Hemisphere are plotted in Figure 4.15. This space represents about one third of the variance in the control integration, while around half of the change due to enhanced greenhouse forcing projects onto it. The PDF from the 1×CO₂ integration (Figure 4.15a) resembles a Gaussian distribution centred on the origin. A separate 1000-year segment of the multiple-millennia 1×CO₂ integration was used here to reduce the bias when projecting onto the EOFs of the control segment of the integration. There is no clear evidence of multi-modality, suggesting that these modes are not related to fluctuations between regimes on the time scales examined here.

The changes in the warmer climates (Figures 4.15b and c) are dominated by a translation of the PDF, with no obvious changes occurring in its shape. The PDFs from the transient integrations (Figures 4.15d and e) indicate that this translation is linear. Consistent with this, the annual mean state, represented in Figure 4.15 by the coloured dots, progresses gradually from the 1×CO₂ state to the warmer states. Thus, in the Northern Hemisphere the projection of change in physical SLP appears overwhelmingly linear on interannual time scales.

4.6.2 Extratropical Southern Hemisphere Physical SLP

The estimated PDFs of physical SLP in the extratropical Southern Hemisphere, in the space spanned by the top two PCs of the control integration, are plotted in Figure 4.16. As in the Northern Hemisphere, the climate system is relatively Gaussian in this reduced space, and the equilibrium climate change takes the form of a linear migration of the PDF. However, due to the lag in ocean warming and sea ice retreat, the trends in the transient integrations do not resemble the equilibrium changes, and instead reflecting decreasing pressure over
Figure 4.15: The estimated probability density functions (PDFs) of physical SLP in the space spanned by the top two principal components (PCs) of the $1\times CO_2$ control integration in the extratropical Northern Hemisphere. The PDFs from a) a second $1\times CO_2$, b) the $2\times CO_2$, and c) the $4\times CO_2$ equilibrium integrations and the d) $1-2\times CO_2$ and e) $1-4\times CO_2$ transient integrations are shown. See Section 4.3 of the text for a description of the generation of these estimated PDFs. For a), b) and c) a smoothing parameter of $h = 0.4$ is used, while for d) and e) a value of $h = 0.675$ is used. Contours denote multiples of 0.02, while the units on the axes are the standard deviations of the PCs in the control integration. In d) and e) the coloured dots represent each annual mean realisation of the climate system in this reduced space.
Figure 4.16: As in Figure 4.15, but the estimated PDFs of physical SLP in the space spanned by the top two PCs of the control integration in the extratropical Southern Hemisphere.

Antarctica. The projection onto the SAM-like mode in the transient integrations mimics the projection onto the NAM-like mode in the Northern Hemisphere.

4.6.3 Global Standardised SLP

With global standardised SLP, the climate change projects partially onto the first and third EOFs and negligibly onto the remaining top patterns (Figure 4.11). The estimated PDFs of standardised SLP in the space spanned by the two corresponding PCs of the control integration are plotted in Figure 4.17. As with physical SLP, at $1\times$CO$_2$ the PDF of the climate system is relatively Gaussian, and changes occur primarily through a translation
Figure 4.17: As in Figure 4.15, but the estimated PDFs of standardised SLP in the space spanned by the first and third PCs of the control integration over the entire globe.

in the warmer climates. The change is relatively linear in the transient integrations, progressing gradually from the $1 \times CO_2$ state to the warmer states. This latter point is curious considering that the first EOF represents an SAM-like mode of variability. The representation of this mode in physical SLP in the extratropical Southern Hemisphere (the first EOF) evolves differently in the transient integrations than between the equilibrium integrations. This difference between the standardised and physical versions of this mode must result from the importance of tropical variability in the standardised version.
4.6.4 Extratropical Northern Hemisphere Physical SAT

The leading EOF of physical SAT in the extratropical Northern Hemisphere accounts for only a negligible portion of the change under enhanced greenhouse forcing. The estimated PDFs in the space spanned by the corresponding PC of the control integration (not shown) behave similarly to those examined above; the PDFs resemble a Gaussian distribution whose location shifts linearly, but only slightly, as the climate warms.

4.6.5 Extratropical Southern Hemisphere Physical SAT

The two leading PCs of extratropical Southern Hemisphere physical SAT represent fluctuations between ice-free and ice-covered states with associated changes in ocean convection, resulting in regime behaviour clearly visible in Figure 4.18. Obviously, in the warmer climates both PCs would tend towards the ice-free regimes. Indeed, the PDF from the $4\times CO_2$ integration does not reveal any regime-like behaviour. Meanwhile, the regime behaviour in the $2\times CO_2$ integration results from sea ice-related variations very different from that represented by these two EOFs in the $1\times CO_2$ integrations. Nevertheless, the PCs in the warmer climates have evolved much further than the ice-free regime, reflecting a warming of the ocean. The linear warming dominates the change here. There is a change in the regime behaviour, but it is unclear whether the regimes in the warmer climates are represented in the control climate, since the linear warming shifts the mean state well away from the control climate. Of course, since these modes do not remain important in the warmer climates, this gradual warming cannot be interpreted using the linear perspective.

4.6.6 Global Standardised SAT

As with physical SAT in the Southern Hemisphere, the estimated PDFs of global standardised SAT in the space spanned by the top PCs (not shown) indicate some regime behaviour. However, once again the linear translation of the climate state dwarfs any changes in the
Figure 4.18: As in Figure 4.15, but the estimated PDFs of physical SAT in the space spanned by the top two PCs of the control integration over the extratropical Southern Hemisphere. A smoothing parameter of $h = 0.2$ is used. The first contour denotes the 0.02 level, while the other contours represent multiples of 0.08.
behaviour in the modes. Since the first and third EOFs remain important in the warmer climates, this change could be consistent with the linear perspective. However, as noted in Section 4.5 this change primarily reflects the overall warming trend rather than the pattern of that change, and thus the physical reality of this projection is questionable.

4.6.7 Summary

The results of Section 4.5 demonstrated that in SLP, substantial fractions of the climate change due to enhanced greenhouse forcing project onto the dominant modes of variability. The results of this section indicate that this projection is predominantly linear. Regime-like behaviour was generally not evident in the estimated PDFs from the second 1\times CO_2 integration, and the climate change largely manifested itself as a linear translation away from the 1\times CO_2 state towards states not frequently visited in the 1\times CO_2 climate. Unlike SLP, evidence of regime behaviour was found for SAT modes related to sea ice variability. Furthermore, the climate change was partly expressed as a shift of the climate system towards the ice-free regimes.

4.7 Discussion

Present and future climate change due to increasing anthropogenic emissions of greenhouse gases is now generally accepted (Mitchell et al. 2001, Cubasch et al. 2001); however, the nature of that change is still a topic of considerable debate. Here we have evaluated two interpretations of this climate change, whereby it projects onto the pre-existing low frequency natural modes of variability of the climate system, using integrations of a coupled GCM.

In order to interpret the concept of climate change projecting onto the dominant modes of variability, it must be established whether these modes remain important in different climates. It was found that the modes of SLP and those of SAT unrelated to sea ice variability do indeed remain important. However, modes related to sea ice variability,
which dominate in the Southern Hemisphere, were no longer found to be important in the warmer climates since the sea ice had retreated to different locations. This implies that while the nonlinear perspective of the projection of climate change may apply to all modes, the linear perspective may apply only to the SLP and non-sea ice SAT modes.

In several of the domains examined the climate change was found to project onto some of the dominant modes of variability. In most cases, this projection tended to be strongest onto the more dominant patterns. This suggests that the climate change will tend to project onto only a single mode of variability, rather than a combination of the more dominant modes. The change in SAT projected only partially onto the top modes, but this generally reflected the overall warming rather than the pattern of that warming. The changes in SLP, however, tended to represent the pattern of the change, since the mean changes were of much smaller magnitude.

In the Northern Hemisphere, SLP was found to project only partly onto the top two modes. This result, along with the findings of Fyfe et al. (1999), is at odds with the conclusion of Shindell et al. (1999) that a modelled stratosphere is necessary to simulate a trend in the NAM, since the GFDL model does not resolve the stratosphere. In the Southern Hemisphere, the change projected almost entirely onto the SAM-like mode in the transient integrations, as is found by Fyfe et al. (1999) and Kusher et al. (2001); however, it projected negligibly between the warmer equilibrium integrations and the control simulation. This suggests that the strong projection onto this pattern is only an initial response. An important question is whether the projection onto the SAM, and the much weaker projection onto the NAM, represent only a surface manifestation of a “monsoon”-like response, or whether they are effectively barotropic and extend to the higher regions of the annular modes. While this issue was beyond the scope of the present work, the results of Kushner et al. (2001) suggest that indeed this response extends at least up to the tropopause.

In all SLP domains examined, no regime behaviour was evident in the projection of
the climate state onto the top modes. Furthermore, the climate change manifested itself in a linear translation of these modes, moving far beyond regions of the climate attractor frequented in the control integration. Unlike for SLP, regime behaviour was clearly visible in some of the SAT modes, and the residence frequency of the climate system in these regimes did change with enhanced greenhouse forcing. Nevertheless, these changes accounted for little of the projection of the climate change onto these modes, which instead mostly reflected an overall warming, and thus represented only a small fraction of the overall climate change.

An evident caveat to these results is the question of the applicability of the nonlinear interpretation to annual mean fields. In its original formulation by Palmer (1999), this hypothesis interprets long term climate change in terms of the high frequency (~one week) fluctuations of the atmosphere. It should be possible to extend this theory to the lower frequency interannual variability of the coupled climate system. However, modes operating at higher frequencies than the interannual time scale would not be detected in annual mean data. SLP does not depend as much on oceanic and sea ice variability as does SAT, and notably support for regime behaviour was found only in the SAT domain. This implies that the weak support here for the low frequency nonlinear interpretation does not automatically apply to this interpretation in its original formulation.

This study aimed to test interpretations of climate change projecting onto the dominant modes of variability, based on linear and nonlinear perspectives of the climate system, using integrations of a coupled GCM. Modes of variability of SAT did exhibit behaviour consistent with the nonlinear perspective. While changes in SAT did project partially onto these modes, this tended to reflect the overall warming rather than the pattern of the change. Consequently, changes in regime behaviour related to the retreat of the sea ice edge, consistent with the nonlinear interpretation, only represented a small fraction of the climate change. Changes in SLP in the Southern Hemisphere projected very strongly in the transient integrations onto an SAM-like mode in a manner consistent with the linear interpretation; however, this pattern of change was absent between the equilibrium integrations. Changes
in SLP in other domains only projected partially onto some modes. In general, climate change tended to project most strongly onto the more important modes. These results indicate that climate change may project onto dominant modes of interannual variability in some domains, but not in others, and that this projection will generally be more consistent with the linear perspective on these time scales.
Chapter 5

The effect of ocean mixing parametrisation on the enhanced CO$_2$ response of the Southern Hemisphere mid latitude jet

5.1 Introduction

A curious result in Chapter 4 was the striking difference in the sea level pressure (SLP) response over the Southern Hemisphere between the transient and equilibrium warming scenarios. These differences apparently are a consequence of the very different warming occurring in this hemisphere, and particularly over the Antarctic Ocean. The slow warming in the transient simulations results from vigorous sequestration of heat over the Antarctic Ocean. This implies that models with different representations of ocean processes, which may result in different surface warming responses, could also produce rather different tropospheric responses.

The current generation of climate models typically predict a hemispheric asymmetry in the surface temperature response to rising greenhouse gas concentrations (Cubasch et al. 2001). In particular, while warming tends to increase monotonically with latitude in the Northern Hemisphere, it reaches a minimum in the Southern Hemisphere around 60°S.
This difference is ascribed to deep vertical mixing in the Antarctic Ocean which draws heat down from the surface. Studies comparing the performance of these models against observational measurements suggest that the horizontal/vertical diffusion parametrisation of sub-grid scale mixing used results in an overestimation of the heat uptake at these latitudes (England 1995, Robitaille and Weaver 1995, McDougall et al. 1996). This raises the possibility that this climate change asymmetry is unrealistic. Indeed, recent models using the more physically based isopycnal, eddy mixing parametrisation of Gent and McWilliams (1990) produce a much larger warming over the Antarctic Ocean, resulting in a more hemispherically symmetric pattern (Wiebe and Weaver 1999, Flato and Boer 2001).

Since the tropospheric circulation is driven in part by surface energy fluxes, it stands to reason that the implementation of varying ocean mixing schemes may result in important differences in the simulated response of the Southern Hemisphere atmosphere to anthropogenic forcing. Most of the tropospheric circulation change observed during the past few decades over the Southern Hemisphere consists of a zonally symmetric shift of mass from high to mid latitudes, reflecting a poleward shift of the mid latitude jet (Thompson et al. 2000, Thompson and Solomon 2002). A similar response is found in climate model simulations forced with rising CO$_2$ concentrations (Fyfe et al. 1999, Kushner et al. 2001, Chapter 4). However, all of the models used in these studies implement some version of the horizontal/vertical diffusion scheme, and so it remains unclear how robust the atmospheric response is to the ocean mixing parametrisation.

Here we examine the consequences of the differences in the ocean mixing parametrisation on the mid latitude jet in global warming integrations of two versions of the Canadian Centre for Climate Modelling and Analysis (CCCma) coupled model (Boer et al. 2000a,b, Flato and Boer 2001). This model was described in Section 2.2, with the simulations listed in Table 2.1. Here we use the full ensembles (three members) of each version forced with observed and predicted increases in greenhouse gases and sulphate aerosols. The models use an identical atmospheric component, but differ slightly in the oceanic component. In

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the first version, CGCM1, ocean mixing is parametrised with horizontal/vertical diffusion, while the second version, CGCM2, uses the Gent and McWilliams (1990) parametrisation for mixing associated with mesoscale eddies. Differences also exist in the sea ice schemes and in integration procedures.

Figure 5.1 shows the change in annual mean and zonal mean surface temperature in the ensembles of simulations of these two model versions. The most striking feature in the CGCM1 ensemble is the much smaller surface warming occurring over the Antarctic Ocean (~60°S) than elsewhere. Conversely, in the CGCM2 ensemble there is a much more even warming in the mid and low latitudes, reflecting a much larger warming over the Antarctic Ocean than occurs in CGCM1. The two model versions also use somewhat different sea ice schemes and different procedures for the ocean spin up and calculation of flux adjustments; however, Flato and Boer (2001) ascribe the difference in the response between the two model versions primarily to the difference in ocean mixing schemes rather than to other model differences. Thus while these two model versions are not designed specifically for a sensitivity study, they are nevertheless appropriate for a diagnosis of the effect of ocean mixing parametrisation.
Figure 5.2: 2050-2099 change from the respective control simulation in the annual and zonal mean meridional temperature gradient in the ensembles of a) CGCM1 and b) CGCM2 global warming simulations. Values have been smoothed and are plotted in °C per degree of latitude, with zero change at each pressure level denoted by the dotted lines. The grey shading represents the zonal mean topography. The dark grey line denotes the location of the maximum at each level in the mid latitudes.

5.2 Results

Tropospheric circulation in the mid latitudes is driven in part by gradients in the meridional profile of temperature. Figure 5.2 shows the change in the annual mean and zonal mean meridional temperature gradient at various pressure levels in the troposphere occurring in the ensembles of global warming simulations of CGCM1 and CGCM2. A striking feature in the CGCM1 ensemble is the maximum increase in the surface temperature gradient occurring around 50°S. A considerably smaller maximum occurs in the CGCM2 ensemble. This results from the much smaller warming over the Antarctic Ocean occurring in the CGCM1 simulations. With both model versions this maximum in the increase in the temperature gradient has a signature up through to the mid troposphere, with this feature being more visible in the CGCM1 ensemble due to the less uniform surface warming.

Since the climatological maximum of the meridional temperature gradient at 500 hPa occurs near 50°S (not shown), the effect of the increase near 55°S is to shift this maximum
Figure 5.3: The evolution of the annual and zonal mean position of hemispheric maxima in one of the global warming simulations of a) CGCM1 and b) CGCM2. Anomalies from the means of the respective control simulations are shown. Black: 200 hPa zonal wind. Red: 200 hPa eddy momentum forcing (shifted 2° south, and only available for CGCM1). Blue: 500 hPa meridional temperature gradient (shifted 4° south).

polewards. This shift is visible in Figure 5.3a, which shows the evolution of the latitude of the annual mean and zonal mean maximum gradient at this pressure level in one of the simulations of CGCM1. The maximum shifts about 3° toward the south pole over the 201 years of forced change. Also shown in Figure 5.3a are the latitudes of the maxima of the 200 hPa eddy momentum flux convergence and 200 hPa zonal wind. The former is defined as $-\partial u'v'/\partial y$, where $u$ and $v$ are the zonal and meridional winds, $t$ is time, and $y$ is latitude. The overbars indicate the zonal mean, while primes denote deviations from that mean. The climatological means of all three quantities lie around 50°S. Notably, both the eddy momentum flux convergence and the zonal wind maxima shift about 3° toward the south pole, as with the temperature gradient maximum. The same quantities from one of the CGCM2 simulations are also shown in Figure 5.3, except for the eddy momentum flux convergence which is not available. The poleward shift of these structures is considerably smaller in the CGCM2 simulation than in the CGCM1 simulation, consistent with the smaller increase in the surface temperature gradient in the mid latitudes.

The poleward shift of the mid latitude jet reflects a zonally symmetric transfer of mass from high to mid latitudes, which is noted by Fyfe et al. (1999) in one of the CGCM1 global warming simulations, and in Chapter 4 and by Kushner et al. (2001) in simulations with.
other models. This response pattern closely resembles the positive phase of the Southern Annular Mode (SAM, also known as the Antarctic Oscillation), which is the dominant mode of variability in the Southern Hemisphere troposphere (Nigam 1990, Thompson and Wallace 2000). The implication then is that similar dynamics to those that maintain the SAM fluctuations may be involved in the mean climate response. The SAM fluctuations are driven by the convergence of momentum fluxes from transient eddies (Hartmann and Lo 1998). Lorenz and Hartmann (2001) compare the SAM variability to changes in the meridional temperature gradient, since this is an important factor in the development of eddies (Holton 1979). They find that the position of the eddy momentum flux convergence, and thus of the jet itself, follows fluctuations in the profile of the temperature gradient in the lower troposphere. Poleward (equatorward) shifts of the jet occur when the gradient around 60°S is larger (smaller) than normal, and vice versa around 40°S.

This covariability of the latitudes of the zonal wind, eddy momentum flux convergence, and meridional temperature gradient maxima is visible in Figure 5.3. In the control simulations of the two model versions the correlations between the three quantities are 0.8 or greater. Thus, the three quantities plotted in Figure 5.3 describe not only the mean long term change, but also the SAM variability. This supports the idea that some of the dynamics involved in the conversion of energy from baroclinic instabilities to the mean flow, which are important for SAM variability, also play an important role in the long term forced shift of the jet. An interesting point is that the climate change in the model simulations manifests itself as a mean shift of the circulation, without affecting the variability (Fyfe 2002). The SAM fluctuations of the jet about the changing mean latitude are indistinguishable between different sections of the simulations. Thus while forcing by transient eddies drives both the change and variability, the mean change does not modify the ability of this mechanism to produce SAM mode fluctuations.

Flato and Boer (2001) find that the recent observed surface warming over the Southern Hemisphere is more closely reproduced in the CGCM2 simulations than in those of CGCM1.
Thus this raises the question as to whether the change in atmospheric circulation in the CGCM2 simulations is also closer to the observed change. As stated earlier, recent Southern Hemisphere climate change also consists of a zonally symmetric poleward shift of the mid latitude jet (Thompson et al. 2000, Thompson and Solomon 2002). Figure 5.4 shows the first principal component of the zonal mean geostrophic wind averaged over the year at 200 hPa south of 20°S in the National Centers for Environmental Prediction and National Center for Atmospheric Research (NCEP-NCAR) reanalysis during the years 1979-2001 (Kistler et al. 2001). The geostrophic wind is almost identical to the actual wind at this altitude and is shown due to data availability. This principal component measures fluctuations in the position of the mid latitude jet and thus is an index of both the SAM and the mean climate response. The latter is visible as a positive trend which is significantly different from zero at the 10% level with respect to the variability of the control integrations of both model versions. However, it does not differ significantly from the trends predicted in the global warming simulations of either of the model versions, also shown in Figure 5.4. The responses in the two model versions are only beginning to diverge over this period, so it is difficult to distinguish between them at this early stage. Nevertheless, this comparison indicates that enhanced greenhouse warming is a plausible cause of the observed change in Southern Hemisphere tropospheric circulation.

5.3 Discussion and Conclusion

Modelling studies indicate that use of the Gent and McWilliams (1990) parametrisation of sub-grid scale ocean mixing improves the correspondence of ocean properties between models and observations, both of the climatological mean (England 1995, Robitaille and Weaver 1995, McDougall et al. 1996) and of the long term trend (Wiebe and Weaver 1999, Flato and Boer 2001). Here we have examined the effect of these changes on CO₂ forced trends in the tropospheric circulation and found that the choice of ocean mixing parametrisation can
have substantial effects. In particular, implementation of the Gent and McWilliams (1990) scheme produces a much smaller poleward shift of the mid latitude jet than occurs when the more common horizontal/vertical diffusion scheme is used in the same model.

Recent studies indicate that changes in the stratospheric polar vortex, for instance due to ozone depletion, may propagate down into the troposphere and result in a poleward shift of the mid latitude jet at times when the two circulations are coupled (Sexton 2001, Thompson and Solomon 2002, Gillett et al. 2002, Polvani and Kushner 2002). Furthermore, Kushner et al. (2001) note that imposing an uniform surface warming in the “aquaplanet” model of Lee (1999) produces a similar change. Combined, these results suggest that meridional shifts in the mid latitude jet are a preferred response in the Southern Hemisphere to a variety of external forcings, and thus may not be robust as a “fingerprint” indicator of the nature of the forcing (e.g. natural or anthropogenic).

The analysis shows that the choice of sub-grid ocean mixing parametrisation in a coupled model can have an important effect on changes in tropospheric circulation in the Southern Hemisphere under global warming, as the mixing scheme alters the surface temperature
response. Thus the large poleward shifts of the mid latitude jet found in previous modelling studies (Fyfe et al. 1999, Kushner et al. 2001, Chapter 4) may be overestimates, since the models involved all use simpler mixing schemes which likely overestimate heat uptake by the Antarctic Ocean. Nevertheless, a similar, but smaller, response pattern results when the more realistic Gent and McWilliams (1990) mixing scheme is used, indicating that a poleward shift may still occur in response to anthropogenically forced warming.
Chapter 6

Conclusion

The global mean warming over the past century is well established (Folland et al. 2001) and has been found to result from the anthropogenic emission of greenhouse gases (Cubasch et al. 2001). The predicted future emission of these gases, along with the slow reaction time of the climate system, implies that an even larger warming will occur through the coming century. However, the degree to which past changes in other aspects of the climate system relate to anthropogenic activities has received less attention. Variations in many of these more directly impact human society than do comparable variations in large scale temperature. Consequently, the possibility of changes in these climate properties due to anthropogenic forcing requires further investigation.

The aim of this study is to improve our understanding of how anthropogenic forcing may affect two properties of the climate system, specifically the diurnal temperature range (DTR) over land areas and the behaviour of natural modes of tropospheric variability. The DTR has decreased over global land areas during the last half century at a comparable rate to the mean warming (Easterling et al. 1997, New et al. 2000, Jin and Dickinson 2002), while indices of some modes of atmospheric variability have drifted over the past few decades (e.g. Thompson et al. 2000, Thompson and Solomon 2002). This coincides with an accelerated human-induced warming and so the possibility exists that human activities have affected these two properties.

Direct comparison of an ensemble of simulations of a climate model including observed increases in greenhouse gases and sulphate aerosols revealed a decrease in the DTR, as observed. A main difference though was in magnitude, with the model predicting a trend
only one third as large as has actually occurred. However, the model produced decreases around the globe, consistent with the observations. Nevertheless, the regional patterns of this global trend were not robust in the ensemble of simulations over the period with observational measurements. The DTR decrease continued through the twenty-first century and was strongest in the mid latitudes.

In the model simulations, the variations in the DTR were much more sensitive to changes in feedbacks than in direct forcings. In particular, the DTR was found to be insensitive to the scattering of sunlight by sulphate aerosols and to the increased mean temperature. Instead, variations resulted mostly from changes in clouds and soil moisture. Consequently, the decreasing DTR trends arose from increases in the reflection of solar radiation by increased clouds, moderated by decreases in soil moisture, mostly through its effect on the ground heat capacity. The exception to this relation occurred in the mid latitudes during the winter, when snow cover reduced the influence of these two factors. DTR changes during this season were a consequence of the simple land surface scheme which produced unrealistic behaviour when the mean temperature was near the freezing point.

Two possible interpretations of forced climate change view it as projecting, either linearly or nonlinearly, onto the natural dominant modes of variability of the climate system, and these were evaluated on interannual time scales using simulations of a climate model. The dominant modes of sea level pressure (SLP) remained important in warmer climates, while those of surface air temperature (SAT) were often related to sea ice variations and so disappeared once the ice retreated. In general, climate change tended to project most strongly onto the more dominant modes. The change in SLP projected partially onto the top two modes in the Northern Hemisphere. In the Southern Hemisphere the change projected negligibly onto the dominant patterns between equilibrium climates but very strongly onto the Southern Annular Mode (SAM) pattern in the transient simulations. Changes in SAT projected partially onto the dominant modes but related more to the mean warming rather than the pattern of change. In all SLP domains the projection
of climate change overwhelmingly manifested itself as a linear translation in the modes, consistent with the linear interpretation. In SAT domains related to sea ice variability the projection reflected an increased tendency toward ice-free regimes, consistent with the nonlinear perspective; however this nonlinear projection represented only a small portion of the overall climate change.

The primary difference between the SAT changes in the transient simulations and between equilibrium simulations was a much larger warming over the Antarctic Ocean in the equilibrium response. This suggested that the differing degree to which the SLP response projected onto the SAM pattern depends strongly on the SAT response in this region. Modelling studies using different parametrisation schemes to represent sub-grid scale ocean mixing produce very different surface warming over this area, and the implication is that this could substantially affect the atmospheric circulation response. Indeed, comparison of simulations of two versions of the same model using different ocean mixing schemes produced very different poleward shifts of the mid latitude jet, of which the SAM is a measure. Since implementation of the more realistic Gent and McWilliams (1990) mixing scheme resulted in a smaller shift, the tropospheric global warming response projected in the Southern Hemisphere in Chapter 4 and in previous studies (Fyfe et al. 1999, Kushner et al. 2001) may be overestimates, since the models involved all use simpler parametrisations which likely overestimate heat uptake by the Antarctic Ocean.

Together, these results indicate that the climate response to anthropogenic greenhouse gas emissions is likely to have important manifestations other than the mean warming. In particular, the warming will likely have an important diurnal component. It will also likely result in some large scale shifts in the tropospheric circulation which closely resemble extreme states of pre-existing natural modes of variability of the climate system. However, in model simulations these changes depend strongly on small scale processes, in particular those relating to clouds, the land surface, and ocean mixing. Improvements in the representation of these processes are currently among the most actively pursued goals in climate model
development. Thus more reliable estimates of the importance of the observed changes in the DTR and tropospheric circulation as fingerprints of anthropogenic forcing of climate change can be expected in the near future. At this early stage, however, model results suggest that these recent changes are consistent with enhanced greenhouse warming, and indicate that they are likely to continue into the foreseeable future.
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