Chapter 10: Global Climate Projections

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Executive Summary

The future climate change results assessed in this chapter are based on a hierarchy of models, ranging from atmosphere-ocean general circulation models (AOGCMs), and earth system models of intermediate complexity (EMICs), to simple climate models (SCMs). In general, we assess projections of future climate change on scales from global to hundreds of kilometers. Further assessments of regional and local climate changes are provided in Chapter 11. Due to an unprecedented, joint effort by many modeling groups worldwide, climate change projections are now based on multi-model means, differences between models can be assessed quantitatively, and in some instances, probabilistic estimates of important climate system parameters replace expert judgement. New results qualitatively corroborate those given in the TAR.

Mean Temperature
All models regardless of emission scenario used project increases in temperature continuing over the 21st century, driven mainly by increases in anthropogenic GHGs, with the warming proportional to the forcing. The close agreement of warming for the early 21st century in the three SRES (B1, A1B and A2) scenarios run by the AOGCMs (warming averaged for 2011 to 2030 compared to 1980 to 1999, with a range of only 0.05°C , from 0.64°C to 0.69°C) using the non-mitigation scenarios considered here and with only anthropogenic forcing, indicates this warming rate is consistent with that observed for the past few decades (see Chapter. 3).Possible future variations of natural forcings could expand that range somewhat, but nearly half of the non-mitigated early 21st century climate change arises from warming we are already committed to (0.38°C for early century, 2011–2030, in the multi-model mean). By mid-century, in the AOGCMs the choice of scenario becomes more important for the magnitude of warming, with a range over the three SRES scenarios of 0.45°C from 1.28°C to 1.73°C, and with only about a quarter of that warming due to climate change we are already committed to (0.46°C). By late century, there are clear consequences depending on which scenario is followed, with a range of 1.35°C from 1.77°C to 3.012°C, with only about 15% of that warming coming from climate change we are already committed to (0.55°C). These results are also corroborated by a hierarchy of models and probabilistic projections.

Geographical patterns of projected warming show greatest temperature increases at high northern latitudes and over land (roughly twice the global average temperature increase), with less warming over the southern oceans and North Atlantic, consistent with observations during the latter part of the 20th century (see Chapter 2). The pattern of zonal mean warming in the atmosphere, with a maximum in the upper tropical troposphere and cooling in the stratosphere, becomes established early in the 21st century, while zonal mean warming in the ocean is seen first near the surface and in the northern midlatitudes, with the warming gradually penetrating downward during the course of the 21st century, most evident at high latitudes where vertical mixing is greatest.

Global mean temperature projections by AOGCMs for SRES non-mitigation scenarios with relatively low (B1), medium (A1B) and high (A2) emissions, suggest a warming of 1.5 to 2.2°C, 2.2°C to 3.3°C and 2.7°C to 4.0°C (mean ± 20%), respectively, above present levels (1980–2000) towards the end of this century (2090–2100). The simple model results suggest slightly higher warmings mainly due to the harmonized forcing and consistent 19 AOGCM model tunings considered. Furthermore, if carbon cycle feedback uncertainties are included, they affect in particular the upper end of the uncertainty ranges. The simple model projects warming of 1.5°C to 2.8°C (B1), 2.3°C to 4.1°C (A1B), and 3.0°C to 5.0°C (A2) (mean ± 1std), including carbon cycle feedback uncertainties. For the high emission illustrative SRES scenario A1FI, this warming is projected to be between 3.5°C and 5.8°C in 2100.

An expert assessment based on the available constraints from observations and the strength of known feedbacks simulated in GCMs indicates that the equilibrium warming for doubling carbon dioxide, or "climate sensitivity", is likely to lie in the range 2 to 4.5°C, with a most likely value about 3°C. Climate sensitivity is very unlikely to be less than 1.5°C. For fundamental physical reasons, as well as data limitations, values substantially higher than 4.5°C still cannot be excluded, but agreement with observations and proxy data is generally worse for those high values than for values in the 2 to 4.5°C range.

Temperature extremes
There is a very likely risk of increased temperature extremes with heat waves projected to be more intense, more frequent and longer lasting in a future warmer climate. Cold episodes are projected to significantly

decrease in a future warmer climate with a decrease in diurnal temperature range almost everywhere. Decreases in frost days are shown to occur almost everywhere in the mid and high latitudes, with a comparable increase in growing season length.

Mean Precipitation

The current generation of models show that precipitation generally increases in the tropical precipitation maxima (such as the monsoon regimes) and over the tropical Pacific in particular, with general decreases in the subtropics, and increases at high latitudes as a consequence of a general intensification of the global hydrological cycle. Globally averaged mean water vapour, evaporation and precipitation increase.

Precipitation extremes

Intensity of precipitation events increases (i.e., proportionately more precipitation falls for a given precipitation event), particularly in tropical and high latitude areas that experience increases in mean precipitation. More recent analyses indicate that even in areas where mean precipitation decreases (most subtropical and midlatitude regions), precipitation intensity increases mainly because there are longer periods between rainfall events. There is a tendency for summer drying of the mid-continental areas during summer, indicating a greater risk of droughts in those regions. Precipitation extremes increase more than does the mean in most tropical and mid- and high latitude areas.

Snow and ice

As the climate warms, snow cover and sea ice extent decrease; glaciers and ice caps lose mass (indicating a dominance of summer melting over winter precipitation increases) and contribute to sea level rise as documented for the previous generation of models in the TAR. Sea ice reduces in the 21st century both in the Arctic and Antarctic with a rather large range of model responses. The reduction is accelerated in the Arctic, where some models project sea ice cover to disappear entirely in the high forcing A2 scenario in the latter part of the 21st century. Thawing of the upper layer of permafrost is projected to be as much as 90% in the A2 scenario by the end of the 21st century.

Carbon cycle

There is unanimous agreement amongst the coupled climate-carbon cycle models run so far that future climate change would reduce the efficiency of the Earth system to absorb anthropogenic carbon dioxide. As a result, a growingly large fraction of anthropogenic CO_2 would stay airborne under a warmer climate. For the A2 emission scenario, this positive feedback leads to additional atmospheric CO_2 varying between 20 and 200 ppm by 2100. Atmospheric CO_2 concentration ranges between 730 and 1020 ppm by 2100 compared with the standard value of 830 ppm used in the AR4 models without an interactive carbon cycle, and provides an indication of the uncertainty due to future changes in the carbon cycle. Accounting for this higher CO_2 concentration leads to an upper warming estimate 1.2°C higher by 2100 for the climate-carbon cycle coupled models than for the AOGCM simulations with the A2 scenario where the CO_2 concentration is prescribed. As the climate-carbon cycle feedback reduces the land and ocean uptakes of CO_2, it leads to a reduction of the emissions required to achieve a given atmospheric CO_2 stabilization. The higher the stabilization scenario, the larger the climate change, the larger the impact on the carbon cycle, and hence the larger the emission reduction.

Sea level

Global average sea-level rise with respect to 2000 of 34 ± 25 mm is projected under scenario SRES A1B by 2020, 120 ± 60 mm by 2050 and 290 ± 150 mm by 2100, at a rate of 3.6 ± 2.1 mm yr^{-1} during 2080–2100. For an average model, the scenario spread in sea level rise is only 8 mm at 2020; at 2050 it is 50 mm, and at 2100 it is 200 mm. Thermal expansion is the largest component, contributing 230 ± 100 mm by 2100 under scenario SRES A1B, with contraction of glaciers and ice caps adding 88 ± 68 mm. Since the Antarctic ice sheet will receive increased snowfall without experiencing substantial surface melting, it will contribute negatively to sea level, unless there are larger accelerations in ice flow of the kind presently taking place in some West Antarctic ice streams. Current understanding is insufficient to permit projections to be made of this effect, but we note that the recent dynamical imbalance would have to increase by a factor of ~5 by 2100 to outweigh the mid-range projection for increased snow accumulation under scenario A1B. The Greenland ice sheet is projected to lose mass, because increased melting will exceed increased snowfall; it would contribute more to sea level if there are further accelerations in ice flow. Sea level rise is predicted to have strong geographical variability, of size comparable to its global average, and the patterns from different models are not generally similar, so confident regional projections cannot be made.

El Niño

A majority of models show a mean El Niño-like SST response pattern in the tropical Pacific (average central and eastern equatorial Pacific sea surface temperatures warm more than the western equatorial Pacific) with a corresponding mean eastward shift of precipitation over the tropical Pacific and the weakening of the interannual ENSO-Asian Australian monsoon connection. Future changes of ENSO interannual variability differ from model to model. The large inter-model differences in future changes of El Niño amplitude, and the inherent century-timescale variability of El Niño in the models, preclude a definitive assessment at this time of what the projected changes could be.

Monsoons

An increase of precipitation is projected in the Asian monsoon (along with an increase in interannual season-averaged precipitation variability) and the African monsoon in JJA, as well as the Australian monsoon in DJF in a warmer climate. The monsoonal precipitation in Mexico and Central America is projected to decrease in association with increasing precipitation over the eastern equatorial Pacific through Walker circulation and local Hadley circulation changes. However, the uncertain role of aerosols in general, and carbon aerosols in particular, complicates the nature of future projections of monsoon precipitation, particularly in the Asian monsoon.

Sea level pressure

Sea level pressure generally increases over the subtropics and midlatitudes, and decreases over high latitudes (order several millibars by the end of the 21st century) associated with a poleward expansion and weakening of the Hadley Circulation and a poleward shift of the storm tracks of several degrees latitude with a consequent increase in cyclonic circulation patterns over the high latitude Arctic and Antarctic regions. Thus there is a positive trend of the Northern Annular Mode (NAM) and the closely related North Atlantic Oscillation (NAO) as well as the Southern Annular Mode (SAM). The magnitude of that increase is generally more consistent across models and greater for the SAM, though there is considerable spread among the models for the NAO.

Tropical cyclones (hurricanes and typhoons)

Results from embedded high resolution models, and global models, ranging in resolution from $1° \times 1°$ to 20 km, show increased peak wind intensities and increased mean and peak precipitation intensities in future tropical cyclones, with the possibility of a decrease in the number of relatively weak tropical cyclones, and increased numbers of intense tropical cyclones.

Midlatitude storms

Model projections show fewer midlatitude storms in most regions but more intense storms with associated damaging winds to go along with the poleward shift of the storm tracks as noted above. The increased wind speeds associated with these stronger storms result in more extreme wave heights in those regions.

Atlantic Ocean meridional overturning circulation

Those models reasonably consistent with present day observations project a reduction of the Atlantic Ocean meridional overturning circulation (MOC) of up to 60% by 2100. In spite of a slowdown of the MOC in most models, there is still warming of surface temperatures over the North Atlantic Ocean and Europe due to the much larger radiative effects of the increase of GHGs. Although the MOC weakens in most models run for the three SRES scenarios, none shows a collapse of the MOC by the year 2100. No coupled model simulation of the Atlantic MOC shows a mean increase of the MOC in response to global warming by 2100. It is very unlikely that the MOC will undergo a large abrupt transition during the course of the 21st century. At this stage it is too early to assess the likelihood of a large abrupt change of the MOC beyond the end of the 21st century. In experiments with the low (B1) and medium (A1B) emission scenarios, and for which the atmospheric GHG concentrations are stabilized beyond 2100, the MOC recovers from initial weakening within one to several centuries.

Radiative forcing

The radiative forcing by well-mixed greenhouse gases computed with radiative transfer codes in twenty of the AOGCMs used in the AR4 has been compared against results from benchmark line-by-line models. The forcing over the period 1860 to 2000 agrees to within 0.1 W m^{-2} at the tropopause, but AOGCMs significantly overestimate the surface forcing by about 0.5 W m^{-2} at the surface.

Climate change commitment (temperature and sea level)

Results from the AOGCM multi-model climate change commitment experiments (concentrations stabilized at year 2000 for 20th century commitment, and at 2100 values for B1 and A1B commitment) indicate that at any given point in time we are committed to about another 0.5°C warming over the next 100 years after concentrations of GHGs are stabilized. Most of this warming occurs in the first several decades after stabilization; afterwards the rate of increase greatly reduces. Globally averaged precipitation commitment 100 years after stabilizing GHG concentrations amounts to roughly an additional increase of 1 to 2% compared to the precipitation values at the time of stabilization. If GHG concentrations could be reduced, global temperatures would decrease. EMICs with coupled carbon cycle mode components show that for a reduction to zero emissions at year 2100 the climate will take of the order of a thousand years to stabilize, and at that time the temperature and sea level will remain well above their pre-industrial values.

With concentrations stabilised at 2100, sea level rise due to thermal expansion in the 22nd and 23rd centuries is nonetheless greater than in the 21st century in most models, with a commitment of a few 0.1 m per century, reducing over many centuries to reach an eventual level of 0.3–0.6 m per degree of global warming. Under sustained elevated temperatures, some glacier volume may persist at high altitude, but most could be disappear over centuries.

With global warming maintained above 3.1 ± 1.6°C relative to pre-industrial, a level likely to be reached by 2100 under scenario SRES A1B, the Greenland ice sheet would be eliminated, except for remnant glaciers in the mountains, raising sea-level by about 7 m, initially at a rate of up to 0.4 m per century for a global warming of 3°C relative to present-day. The rate of melting would be increased if ice-flow was accelerated by lubrication due to surface meltwater. If the ice sheet were removed, there is medium likelihood that it could not be regenerated even if the climate were subsequently returned to pre-industrial.

GCMs indicate that greater accumulation of snow on the Antarctic ice sheet will cause a negative contribution to sea level, of 0.1 ± 0.1 m per century for a global warming of 3°C relative to present-day. However, this could be counteracted by increased ice discharge into the ocean, especially if the major ice shelves were weakened. In the absence of models for the relevant processes, there is little agreement about what dynamical changes could occur. By analogy with past climate changes, sustained global warming of 2°C has been suggested as a threshold beyond which there will be a commitment to large sea-level contribution from the WAIS, with maximum rates of several mm yr^{-1}. Alternatively, rapid discharge may be transient and insufficient to outweigh the increased snow accumulation.

10.1 Introduction

This chapter addresses various aspects regarding projections of future climate change. Similar chapters have appeared in every IPCC Assessment, so it will be useful at the outset to provide an indication concerning what is new in this chapter since the TAR.

The global coupled climate modeling community has undertaken the largest coordinated global coupled climate model experiment ever attempted to provide the most comprehensive multi-model perspective on climate change of any IPCC assessment. This open process involves experiments with idealized climate change scenarios (i.e., 1% per year CO_2 increase, also included in the newer Coupled Model Intercomparison Project phase 2 and phase 2+ (CMIP2 and CMIP2+) (e.g., Covey et al., 2003; Meehl et al., 2005b), equilibrium $2 \times CO_2$ experiments with atmospheric models coupled to non-dynamic slab oceans, and idealized stabilized climate change experiments at $2 \times CO_2$ and $4 \times CO_2$ in the 1% CO_2 increase simulations. In the idealized 1% per year CO_2 increase experiments, there is no actual real year time line for this type of experiment. Thus, the rate of climate change is not the issue in these experiments, but what is studied are the types of climate changes that occur at the time of doubling or quadrupling of CO_2. Simulations of 20th century climate have been completed that include time-evolving natural and anthropogenic forcings. For future climate change in the 21st century, a subset of three SRES scenario simulations have been selected from the commonly used six marker scenarios (Nakicenovic and Swart, 2000). This subset (A1B, B1 and A2) constitutes a "low", "medium", and "high" scenario among the marker scenarios, and this choice is solely made by the constraints of available computer resources that did not allow for the calculation of all six scenarios. This choice, therefore, does not imply a qualification of, or preference over, the six marker scenarios. By the same argument, it is not within the remit of this report to assess the realism and likelihood of emission scenarios.

Results for the SRES scenarios shown in this chapter begin in the year 2000, and this report is being produced in the 2006 time frame. However, as discussed in this chapter, the forcing over this six year period differs only slightly among the scenarios, with climate change commitment being the main driver on those short timescales. Thus, the forcing and thus the warming for the 1990–2000 time period in the models forced by the SRES scenarios closely resembles the warming observed over that time period (van Vuuren and O'Neill, 2006). Additionally, the warming for the 2000–2006 time period in the models forced by the SRES scenarios when the integrations begin at the year 2000 closely resembles the warming observed over that time period.

The simulations with the subset A1B, B1, and A2 have been performed to the year 2100. Three different stabilization scenarios have been run, the first with all atmospheric constituents fixed at year 2000 values and the models run for an additional 100 years, and the second and third with constituents fixed at year 2100 values for A1B and B1, respectively, for another 100 to 200 years. Consequently, the concept of climate change commitment (for details and definitions see Section 10.7) will be addressed in much wider scope and greater detail than in any previous IPCC assessment. Results based on this AOGCM multi-model data set will be featured in Section 10.3.

Many of the figures in Chapter 10 are based on the mean and spread of the multi model ensemble of comprehensive AOGCMs. The reason to focus on the multi-model mean is that averages across structurally different models empirically show better large-scale agreement with observations, because individual model biases tend to cancel (see Chapter 8). Even though the ability to simulate present day mean climate and variability, as well as observed trends, differs across models, all submitted models are weighted equally in the mean. Since the ensemble is strictly an 'ensemble of opportunity', without sampling protocol, the spread of model is unable to span the full possible range of uncertainty, and a statistical interpretation of the model spread is therefore problematic. However, attempts are made to also quantify uncertainty throughout the chapter based on various other lines of evidence, including perturbed physics ensembles specifically designed to study uncertainty within one model framework, and Bayesian methods using observational constraints. Uncertainties derived from the spread of the multi-model ensemble are generally found to be consistent with estimates from other methods.

In addition to this coordinated international multi-model experiment, a number of entirely new types of experiments have been performed since the TAR to quantify uncertainty regarding climate model response to external forcings. The extent to which uncertainties in parameterizations translate into the uncertainty in climate change projection is addressed in much greater detail. New calculations of future climate change from the larger suite of SRES scenarios with simple models and earth system models of intermediate complexity (EMICs) provide additional information regarding uncertainty related to the choice of scenario. Such models also provide estimates of long-term evolution of global mean temperature, ocean heat uptake, and sea level rise beyond the 21st century, and thus allow us to better constrain climate change commitments.

Another important source of uncertainty concerns the formulation of, and interaction with, the carbon cycle in climate models. Since TAR, much progress has been made in global coupled climate-carbon cycle models. These models allow the prescription of CO_2 emissions, rather than concentrations, and calculate the CO_2-related radiative forcing consistently from the response of the carbon cycle components (ocean, terrestrial and marine biospheres) to the climate changes.

Climate sensitivity has always been a focus in the IPCC assessments, and here we assess more quantitative estimates of equilibrium climate sensitivity and transient climate response (TCR) in terms of not only ranges but also probabilities within these ranges. Some of these probabilities are now derived from ensemble simulations subject to various observational constraints, and no longer rely solely on expert judgement. This gives us a much more complete understanding of model response uncertainties from these sources than ever before. These are now standard benchmark calculations with the global coupled climate models, and are useful to assess model response in the subsequent time-evolving climate change scenario experiments.

With regard to these time-evolving experiments simulating 21st century climate, since the TAR we have seen increased computing capabilities that now allow routine performance of multi-member ensembles in climate change scenario experiments with global coupled climate models. This provides us with the capability to analyze more multi-model results and multi-member ensembles, and yields more probabilistic estimates of time-evolving climate change in the 21st century.

Finally, while future changes in some weather and climate extremes (e.g., heat waves) were addressed in the TAR, there were relatively few studies on this topic available for assessment at that time. Since then, more analyses have been performed regarding possible future changes in a variety of extremes. It is now possible to assess, for the first time, multi-model ensemble results for certain types of extreme events (e.g., heat waves, frost days, etc.). These new studies provide a more complete range of results for assessment regarding possible future changes in these important phenomena with their notable impacts on human societies and ecosystems. A synthesis of results from studies of extremes from observations and model is given in Chapter 11.

Uncertainty of climate change projections has always been a subject of previous IPCC assessments, and there has been considerable new work done on this topic that will be assessed in this chapter. In particular, it is important to keep in mind the sources and propagation of uncertainty in climate model projections (Figure 10.1.1). First, there are multiple emission scenarios for the 21st century, and even at this first stage there is uncertainty with regards to what will be the future time-evolution of emissions of various forcing agents such as greenhouse gases (GHGs) (box at left in Figure 10.1.1). Then these emissions must be converted to concentrations of constituents in the atmosphere. Gas cycle models must be employed, and these models include their own set of parameterisations, assumptions and caveats. Then the concentrations in the atmospheric models produce radiative forcing that acts on the climate system within the atmospheric model components, each with their own radiation schemes and other formulations that affect radiative forcing. Finally, the modelled coupled climate system takes those radiative forcings and produces a future simulated climate. The components of the atmosphere, ocean, sea ice and land surface in each model interact with their sets of strengths and weaknesses to produce a spread of outcomes for future climate. Thus at every step in this process, there are uncertainties and assumptions that must be made to proceed from emissions, to concentrations, to radiative forcing, and eventually to simulated climate changes and impacts.

[INSERT FIGURE 10.1.1 HERE]

This apparent bewildering array of uncertainty suggests that it is difficult to be able to come to any conclusions regarding possible future climate change. However, the use of multi-model ensembles has been shown in other modelling applications to produce simulated climate features that are improved over single models alone (see discussion of Chapters 8 and 9). The expanded use of multi-model ensembles for future climate change therefore provides higher quality and more quantitative climate change information compared to the TAR. The use of large (order 20) global coupled climate multi-model ensembles provides the ability to better quantify differences of model response. A hierarchy of models ranging from simple to intermediate to complex allows better quantification of the consequences of various parameterisations and formulations. Very large ensembles (order hundreds) with single models provide the means to quantify parameterisation uncertainty. Finally, being able to constrain future climate model projections with information from climate characteristics we have already observed, helps us better quantify possible future climate changes.

10.2 Projected Changes in Radiative Forcing

The global projections discussed in this chapter are extensions of the simulations of the observational record discussed in Chapter 9. The simulations of the 19th and 20th centuries are based upon changes in long-lived greenhouse gases (LLGHGs) that are reasonably constrained by the observational record. However, estimates of future concentrations of LLGHGs and other radiatively active species are clearly subject to significant uncertainties. The evolution of these species is governed by a variety of factors that are difficult to predict, including changes in population, energy use, energy sources, and emissions. For these reasons, a range of projections for future climate change has been conducted using coupled AOGCMs. The future concentrations of LLGHGs and the anthropogenic emissions of SO_2, a chemical precursor of sulfate aerosol, are obtained from several scenarios considered representative of low, medium, and high emission trajectories. These basic scenarios and other forcing agents included in the AOGCM projections are discussed in Section 10.2.1. The correspondence between recent trends in emissions and the scenarios and the implications for radiative forcing for the early 21st century are considered in Section 10.2.2.

10.2.1 Radiative Forcing of the Multi-Model Climate Projections

The temporal evolution of the LLGHGs, aerosols, and other forcing agents are described in Sections 10.2.1.1 and 10.2.1.2. Typically, the future projections are based upon initial conditions extracted from the end of the simulations of the 20th century. Therefore, the radiative forcing at the beginning of the model projections should be approximately equal to the radiative forcing for present-day concentrations relative to pre-industrial conditions. The relationship between the modelled radiative forcing for the year 2000 and the estimates derived in Chapter 2 is evaluated in Section 10.2.1.3. Estimates of the radiative forcing in the multi-model integrations for one of the standard scenarios are also presented in this section. Possible explanations for the range of radiative forcings predicted for 2100 are discussed in Section 10.2.1.4, including evidence for systematic errors in the formulations of radiative transfer used in AOGCMs. Possible implications of these findings for the range of global temperature change and other climate responses are summarized in Section 10.2.1.5.

10.2.1.1 The SRES scenarios

The future projections discussed in this chapter are based upon three standard SRES scenarios (Nakicenovic and Swart, 2000) and a new idealized scenario designed to quantify committed climate change due to historical emissions. The models have been integrated to year 2100 using the predicted concentrations of LLGHGs and emissions of SO_2 specified by the A1B, B1, and A2 emissions scenarios. Some of the AOGCMs do not include sulfur chemistry, and the simulations from these models are based upon concentrations of sulfate aerosols from Boucher and Pham (2002) (see Section 10.2.1.2). The simulations for the three scenarios were continued for another 100 to 200 years with all anthropogenic forcing agents held fixed at values applicable to the year 2100. In the commitment simulations, the models have been integrated to 2100 with all anthropogenic forcing agents held fixed at values applicable to the year 2000.

10.2.1.2 Forcing by additional species and mechanisms

The forcing agents applied to each AOGCM used to make climate projections are summarized in Table 10.2.1. The radiatively active species specified by the SRES scenarios are CO_2, CH_4, N_2O, CFCs, and SO_2, which is listed in its aerosol form as SO_4 in the table. The inclusion, magnitude, and time evolution of the remaining forcing agents listed in Table 10.2.1 have been left to the discretion of the individual modelling groups. These agents include tropospheric and stratospheric ozone, all of the non-sulfate aerosols, the indirect effects of aerosols on cloud albedo and lifetime, the effects of land use, and solar variability.

Table10.2.1. Radiative forcing agents in the multi-model global climate projections. The entries have the following meaning: "YES" indicates that the time-varying forcing for the corresponding forging agent is included; "Const." means the forcing agent is set to a constant or annually cyclic distribution. "--" denotes the forcing agent is omitted from the simulations of the 20th century and of the future.

Model	CO2	CH4	N2O	Strat O3	Trop O3	CFCs	SO4	Urban	BC	OC	Nitrate	1st Indirect	2nd Indirect	Dust	Volcanic	Sea Salt	Land Use	Solar
BCCR-BCM2.0	ENSEMBLE	ENSEMBLE	ENSEMBLE			ENSEMBLE	BP	Const.	--	--	--	--	--	Const.	--	Const.	--	--
BCC-CM1	COVEY	COVEY	COVEY	Const.	Const.	COVEY	COVEY	--	--	--	--	--	--	--	--	--	--	--
CCSM3	COVEY	COVEY	COVEY	NCAR	NCAR	COVEY	CIESIN	--	NCAR	NCAR	--	--	--	Const.	Const.	Const.	--	Const.
CGCM3.1(T47)																		
CGCM3.1(T63)																		
CNRM-CM3	ENSEMBLE	ENSEMBLE	ENSEMBLE			ENSEMBLE	BP	Const.	--	--	--	--	--	Const.	--	Const.	--	--
CSIRO-Mk3.0								--	--	--	--	--	--	--	--	--		
ECHAM5/MPI-O	YES	YES	YES	YES	Const.	YES	BP	--	--	--	--	YES	--	--	--	--	--	--
ECHO-G	YES	YES	YES	YES	Const.	YES	YES	--	--	--	--	YES	YES	--	Const.	--	--	Const.
FGOALS-g1.0	COVEY	COVEY	COVEY	Const.	Const.	COVEY	COVEY	--	--	--	--	--	--	--	--	--	--	--
GFDL-CM2.0	YES	YES	YES	YES	YES	YES	YES	--	YES	YES	--	--	--	Const.	Const.	YES	Const.	Const.
GFDL-CM2.1	YES	YES	YES	YES	YES	YES	YES	--	YES	YES	--	--	--	Const.	Const.	YES	Const.	Const.
GISS-AOM	GISS	GISS	GISS			GISS	BP	--	--	--	--	--	--	--	--	--	--	--
GISS-EH	YES	YES	YES	YES	YES	YES	YES	--	YES	YES	YES	--	YES	Const.	YES	Const.	YES	YES
GISS-ER	YES	YES	YES	YES	YES	YES	YES	--	YES	YES	YES	--	YES	Const.	YES	Const.	YES	YES
INM-CM3.0	COVEY	COVEY	COVEY	Const.	Const.	--	COVEY	--	--	--	--	--	--	--	Const.	--	--	Const.
IPSL-CM4	ENSEMBLE	ENSEMBLE	ENSEMBLE			ENSEMBLE	BP	--	--	--	--	YES	--	--	--	--	--	--
MIROC3.2(H)	YES	YES	YES			YES	YES	--	YES	YES	--	YES	YES	YES	Const.	YES	Const.	Const.
MIROC3.2(M)	YES	YES	YES			YES	YES	--	YES	YES	--	YES	YES	YES	Const.	YES	Const.	Const.
MRI-CGCM2.3.2	MRI	MRI	MRI	Const.	Const.	MRI	MRI	--	--	--	--	--	--	--	Const.	--	--	Const.
PCM												--	--					
UKMO-HadCM3	YES	YES	YES	YES	Yes	YES	YES	--	--	--	--	YES	--	--	--	--		
UKMO-HadGEM	YES	YES	YES	YES	YES	YES	YES	--	YES	YES	--	YES	YES	--	Const.	YES	Const.	Const.

The scope of the treatments of aerosol effects in AOGCMs has increased markedly since the TAR. Nine of the AOGCMs include the first indirect effects and six include the second indirect effects of aerosols on cloud properties. Under the more emissions intensive scenarios considered in this chapter, the magnitude of the first indirect (Twomey) effect can saturate. Johns et al. (2003) parameterize the first indirect effect of anthropogenic emissions as perturbations to the effective radii of cloud drops in simulations of the B1, B2, A2, and A1FI scenarios using HadCM3. At 2100, the first indirect forcings range from –0.50 to –0.79 W m^{-2}. The normalized indirect forcing decreases by a factor of 4 from approximately –7 W/mg[S] in 1860 to between –1 to –2 W/mg[S] by the year 2100. Boucher and Pham (2002) and Pham et al. (2005) find a comparable decrease in forcing efficiency of the indirect effect from –9.6 W/mg[S] in 1860 to between –2.1 and –4.4 W/mg[S] in 2100. Johns et al. (2003) and Pham et al. (2005) attribute the decline to the decreased sensitivity of clouds to greater sulphate concentrations at sufficiently large aerosol burdens.

10.2.1.3 Comparison of modelled forcings to estimates in Chapter 2

The forcings used to generate climate projections for the standard SRES scenarios are not necessarily uniform across the multi-model ensemble. Differences among models may be caused by different projections for radiatively active species (see Section 10.2.1.2) and by differences in the formulation of radiative transfer (see Section 10.2.1.4). The AOGCMs in the ensemble include many species which are not specified or constrained by the SRES scenarios, including ozone, tropospheric non-sulphate aerosols, and stratospheric volcanic aerosols. Other types of forcing which vary across the ensemble include solar variability, the indirect effects of aerosols on clouds, and the effects of land-use change on land-surface albedo and other land-surface properties (Table 10.2.1). While the time series of well-mixed greenhouse gases for the future scenarios are identical across the ensemble, the concentrations of these gases in the 19th and early 20th centuries are left to the discretion of individual modelling groups. The differences in radiatively active species and the formulation of radiative transfer affect both the simulations of the 19th and 20th centuries and the scenario integrations initiated from these historical simulations. The resulting differences in the forcing complicate the separation of forcing and response across the multi-model ensemble. These differences can be quantified by comparing the range of forcings across the multi-model ensemble against standard estimates of radiative forcing over the historical record.

The longwave radiative forcings for the SRES A1B scenario from fourteen climate model simulations are compared against estimates using the IPCC TAR formulae (see Chapter 2) in Figure 10.2.1 (panel A). The graph shows the IPCC TAR estimate for the forcing between 1850 to 2000 and the model forcings between the start of the model integrations and 2000. The forcings from the models are diagnosed from changes in top-of-atmosphere fluxes and the forcing for doubling carbon dioxide (Forster and Taylor, 2006). The IPCC TAR and median model estimates of the longwave forcing are in very good agreement over the 21st century, with differences ranging from –0.27 to 0 W m^{-2}. However, the range of the models for the period 2080–2099 is nearly 4 W m^{-2}, or approximately 60% of the median longwave forcing for that time period. For the year 2000, the IPCC TAR and median model values differ by only –0.06 W m^{-2}.

[INSERT FIGURE 10.2.1 HERE]

The corresponding time series of shortwave forcings for the SRES A1B scenario are plotted in Figure 10.2.1 (panel B). It is evident that the differences among the models and between the models and the IPCC estimates are larger for the shortwave band. The IPCC TAR value is larger than the median model forcing by 0.2 to 0.5 W m^{-2} for individual 20-year segments of the integrations. For the year 2000, the IPCC TAR estimate is larger by 0.44 W m^{-2}. In addition, the range of modelled forcings is sufficiently large that it includes positive and negative values for every 20-year period. For the year 2000, the shortwave forcing from individual AOGCMs ranges from approximately –1.5 W m^{-2} to +1.5 W m^{-2}. The reasons for this large range include the variety of the aerosol treatments and parameterizations for the indirect effects of aerosols in the multi-model ensemble.

Since the large range in both longwave and shortwave forcings may be caused by a variety of factors, it is useful to determine the range caused just by differences in model formulation for a given (identical) change in radiatively active species. A standard metric is the global-mean, annually averaged all-sky forcing at the tropopause for doubling carbon dioxide. Estimates of this forcing for nine of the models in the ensemble are given in Table 10.2.2. The shortwave forcing is caused by absorption in the near-infrared bands of CO_2. The range in the longwave forcing is 0.84 W m^{-2}, and the coefficient of variation, or ratio of the standard deviation to mean forcing, is 0.08. These results suggest that up to 21% of the range in longwave forcing in the ensemble for the period 2080–2099 is due to the spread in forcing from the increase in CO_2. The findings also imply that it is not appropriate to use a single forcing value for doubling CO_2 to evaluate the climate sensitivity across the multi-model ensemble. Although the shortwave forcing has a coefficient of variation in excess of 2, the range across the ensemble explains less than 13% of the range in shortwave forcing at the end of the 21st-century simulations. This suggests that species and forcing agents other than carbon dioxide cause the large variation among modelled shortwave forcings.

Table10.2.2. All-sky adjusted forcing for doubling carbon dioxide

Group	**Model**	**Longwave (W m^{-2})**	**Shortwave (W m^{-2})**
CCCma	CGCM 3.1	3.39	–0.07
GISS	GISS-ER	4.21	–0.15
IPSL	IPSL-CM4	3.50	–0.02
CCSR	MIROC 3.2-hires	3.59	—
CCSR	MIROC 3.2-medres	3.66	—
MPI	ECHAM5/MPI-OM	3.98	0.03
NCAR/CRIEPI	CCSM3	4.23	–0.28
UKMO	UKMO-HadCM3	4.03	–0.17
UKMO	UKMO-HadGEM1	4.02	–0.19
Mean ± std. deviation		3.85 ± 0.31	–0.12 ± 0.11

10.2.1.4 Results from RTMIP: Implications for fidelity of forcing projections

To help understand the response of the ensemble of coupled climate models to the various emissions scenarios, the AOGCM modeling community has engaged in a radiative-transfer model intercomparison, or RTMIP (Collins et al., 2006b). The primary objective is to determine the differences in forcing caused by the use of different radiation codes in the AOGCMs used for climate change simulations in the IPCC AR4. The basis of RTMIP is an evaluation of the forcings computed by AOGCMs using benchmark line-by-line (LBL) radiative transfer codes. The comparison is focused on forcing by the LLGHGs CO_2, CH_4, N_2O, CFC-11, CFC-12, and the increased H_2O expected in warmer climates. The data requested for RTMIP is the instantaneous changes in clear-sky fluxes and heating rates. While the relevant quantity for climate change is all-sky forcing, the introduction of clouds would greatly complicate the intercomparison exercise. In addition, the calculations omit the effects of stratospheric thermal adjustment to forcing using fixed dynamical heating (FDH). This omission facilitates comparison of fluxes from LBL codes and AOGCM parameterizations. It should be noted that our omission of FDH means that the results of this intercomparison are not directly comparable to the estimates of forcing at the tropopause in the IPCC TAR, since the latter include the effects of adjustment. The effects of adjustment on forcing are approximately –2% for CH_4, –4% for N_2O, +5% for CFC-11, +8% for CFC-12, and –13% for CO_2 (IPCC, 1995; Hansen et al., 1997).

The results include numerical calculations from fourteen AOGCM groups representing twenty of the models in the multi-model ensemble. The benchmark calculations represent a synthesis of five different LBL codes. All the calculations are for climatological mid-latitude summer conditions. The sets of calculations in RTMIP cover several forcing configurations, including forcing by greater amounts of CO_2, forcing by increased concentrations of other LLGHGs, and the effect of more water vapour in response to forcing by CO_2. The total (longwave plus shortwave) forcings at 200mb, a surrogate for the tropopause, are shown in Table 10.2.3. For example, the differences between the fluxes calculated for 3b and 3a represent the instantaneous clear-sky radiative forcing from changes in LLGHGs between 1860 and 2000 under summer mid-latitude conditions.

Total forcings calculated from the AOGCM and LBL codes due to the increase in LLGHGs from 1860 to 2000 differ by less than 0.04, 0.49, and 0.08 Wm^{-2} at the top of model, surface, and pseudo-tropopause at 200mb, respectively. (Table 10.2.3). Based upon the Student t-test, none of the differences in mean forcings shown in Table 10.2.3 are statistically significant at the 0.01 level. This indicates that the ensemble-mean forcings are in reasonable agreement with the LBL codes. However, the forcings from individual models, for example from doubling CO_2, span a range at least 10 times larger than that exhibited by the LBL models.

Table 10.2.3. Total instantaneous forcing at 200 hPa (W m^{-2}) from AOGCMs and LBL codes in RTMIP

Radiative Species	CO_2	CO_2	N_2O+ CFCs	CH_4+ CFCs	All LLGHGs	H_2O
Forcing[a]	2000–1860	2×–1×	2000-1860	2000–1860	2000–1860	1.2×–1×
<AOGCM>[b]	1.56	4.28	0.47	0.95	2.68	4.82
σ(AOGCM)[b]	0.23	0.66	0.15	0.30	0.30	0.34
<LBL>	1.69	4.75	0.38	0.73	2.58	5.08
σ(LBL)	0.02	0.04	0.12	0.12	0.11	0.16

Notes:
(a) 2000-1860 is the forcing due to an increase in the concentrations of radiative species between 1860 and 2000. 2×-1× and 1.2×–1× are forcings from increases in radiative species by 100% and 20% relative to 1860 concentrations.
(b) <M> and σ(M) are the mean and standard deviation of forcings computed from model type M (AOGCM or LBL).

The forcings from doubling CO_2 from its concentration at 1860 AD are shown in Figure 10.2.2 (panel A) at the top of the model (TOM), 200 hPa, and the surface. The sums of the longwave and longwave forcings at Figure 10.2.2 correspond to the total forcings given in Table 10.2.3. The AOGCMs tend to underestimate the longwave forcing at these three levels. The relative differences in the mean forcings are less than 8% for the pseudo-tropopause at 200 hPa but increase to approximately 13% at the TOM and to 33% at the surface. The small errors at 200 hPa may reflect earlier efforts to improve the accuracy of AOGCM calculations near the tropopause in order to produce reasonable estimates of radiative forcing. In general, the mean shortwave forcings from the LBL and AOGCM codes are in good agreement at all three surfaces. However, the range in shortwave forcing at the surface from individual AOGCMs is quite large. The coefficient of variation (the ratio of the standard deviation to the mean) for the surface shortwave forcing from AOGCMs is 0.95. In response to a doubling in CO_2, the relative humidity increases by approximately 20% through much of the troposphere. The changes in shortwave and longwave fluxes due to a 20% increase in water vapour in are illustrated in Figure 10.2.2 (panel B). The mean longwave forcing from increasing H_2O is quite well simulated with the AOGCM codes. In the shortwave, the only significant difference between the AOGCM and LBL calculations occurs at the surface, where the AOGCMs tend to underestimate the magnitude of the reduction in insolation.

[INSERT FIGURE 10.2.2 HERE]

Other calculations show that a few of the participating AOGCMs do not include the effects of CFCs on the longwave fluxes. In addition, all AOGCMs omit the effects of CH_4 and N_2O on the shortwave fluxes despite the large magnitude of the surface forcings by these gases. While the omission of N_2O does not introduce a large absolute error in the forcings, the omission of CH_4 introduces an error in the net surface shortwave forcing of 0.5 W m^{-2} relative to the LBL calculations. The biases in the AOGCM forcings are generally largest at the surface level. For five out of seven surface shortwave forcings and four out of seven surface longwave forcings examined in the intercomparison, the probability that the mean AOGCM and LBL values agree, is less than 0.01. The largest biases in the shortwave and longwave forcings from all seven experiments occur at the surface layer.

10.2.1.5 Implications for range in climate response

The results from RTMIP imply that a fraction of the spread in climate response discussed in this chapter is due to diverse formulations of radiative transfer rather than differences in forcings or feedbacks among the models. Many of the climate responses (e.g., global mean temperature) scale linearly with the radiative forcing to first approximation. Therefore, systematic errors in the calculations of radiative forcing should produce a corresponding range in climate responses. Assuming that the RTMIP results (Table 10.2.3) are globally applicable, the range of forcings for 1860 to 2000 in the AOGCMs should introduce a ±11% relative range for 2000 in the responses that scale with forcing. The corresponding relative range for doubling CO_2, which is comparable to the change in CO_2 in the B1 scenario, is ±15%.

10.2.2. Recent Developments in Forcing Projections for the 21st Century

The SRES scenarios used as the basis for the AR4 climate projections are constructed using models that depend on predictions regarding socioeconomic factors, energy use, and emissions. The consistency of the scenarios with observations of atmospheric concentrations and with new societal data collected since the construction of the SRES scenarios is discussed in Section 10.2.2.1. The implications of these findings for future radiative forcing are also assessed. The impact of radiative species omitted from the SRES specifications are evaluated in Section 10.2.2.2. These additional species contribute to the range of forcings shown in Figure 10.2.1. It is likely that these species and forcing agents will be incorporated in an increasing number of AOGCMs used in upcoming international climate assessments.

10.2.2.1 Projections for radiative species considered in SRES

Recent trends in emissions are generally consistent with the range encompassed by the marker SRES scenarios (van Vuuren and O'Neill, 2006). Emissions growth for CO_2 projected by the SRES scenarios between 1990 and 2000 is somewhat larger than current emissions inventories, but the total emissions for 2000 are in good agreement. Perhaps the largest discrepancies between the SRES scenarios and recent data are related to the decline in SO_2 emissions. The decline of 3% predicted by SRES is considerably smaller in magnitude than the 20% decline that actually occurred. The differences in CO_2 and SO_2 are traceable to trends in China and the Reforming Economies that diverge from the assumptions used to construct the SRES projections. Given the central role of China in these discrepancies, its emissions are examined in greater detail below.

Recent trends in some of the long-lived greenhouse gases (LLGHGs) differ from the trends in the early 21st century in the more emissions-intensive SRES scenarios. In recent projections of ozone and methane for the period 2000 to 2030 (see Section 10.4.2), the B2 scenario is considered a plausible but pessimistic trajectory for methane emissions (Dentener et al., 2005). Recent trends have been consistent with projections assuming maximum feasible reductions (MFR) in which global emissions of CH_4 would be reduced by 110 Tg yr^{-1} by 2030. In the MFR simulations, the radiative forcing by methane between 2000 and 2030 is less than 0.004 W m^{-2}.

Evidence from recent economic activity suggests that emissions of CO_2, CH_4, black carbon (BC), and SO_2 from Asia decreased from 1996 (the peak year) to 2000 (Streets et al., 2001). SO_2 is one of the primary chemical precursors of anthropogenic sulphate (SO_4) aerosol. The amount of CO_2 emitted from all sources fell from 3470 Tg/yr to 3220 Tg/yr during that period, a 7.3% decline. Decreases of 32% in BC emissions and 21% in SO_2 from 1996 through 2000 have been qualitatively confirmed by aerosol measurements in the Asian outflow from Midway Island (Prospero et al., 2003). In situ data on sulphate and nitrate from Midway show that concentrations nearly doubled from 1981 to roughly 1995, and then began to decline. An economic recession in 1997–1998 contributed to a decline in emissions from many countries in east and southeast Asia (Carmichael et al., 2002). In addition, Asia took several steps to reduce air pollution, including closure of some high-sulphur coal mines, closure of inefficient industrial plants, and the institution of a SO_2 reduction program for environmentally sensitive regions.

Streets et al. (2001) suggest that anthropogenic greenhouse-gas emissions from China will rebound from the economic downturn and imposition of emission controls of the late 1990s, but that the emissions growth would probably grow much less rapidly than previous projections. Carmichael et al. (2002) estimate that Asian SO_2 emissions will grow from 34.4 Tg yr^{-1} in 2000 to perhaps 40–45 Tg yr^{-1} in 2020, a value considerably lower than previous estimates as high as 80–110 Tg yr^{-1} (Foell et al., 1995). The new estimates are consistent with the SRES B1 scenario (Nakicenovic and Swart, 2000) but are markedly lower than the A1B and A2 scenarios. Since the Asian emissions are the dominant contribution to the total annual flux of SO_2, these results strongly suggest that the future SO_2 emissions in the A1B and A2 scenarios are unrealistically large. Reduction of the emissions in the A1B and A2 scenarios for consistency with current projections would lead to smaller sulphate radiative forcing. Nonetheless, these scenarios form the basis for two of the standard projection experiments considered in this chapter. The overestimation of SO_2 in all the SRES scenarios is already evident at the starting point of the scenario time series in 2000 (Smith et al., 2001). Several simulations with AOGCMs of the transition from the 20th century to the SRES scenarios have used the modern SO_2 emissions data sets and have therefore had to introduce a transition to the higher SRES time series at the year 2000.

10.2.2.2 Projections for radiative species: Extensions beyond SRES

Estimation of ozone forcing for the 21st century is complicated by the short chemical lifetime of ozone compared to atmospheric transport timescales and by the sensitivity of the radiative forcing to the vertical distribution of ozone. Gauss et al. (2003) have calculated the forcing by anthropogenic increases of tropospheric ozone through 2100 from eleven different chemical transport models integrated with the SRES A2p scenario. The A2p scenario is the preliminary version of the marker A2 scenario and has nearly identical time series of well-mixed greenhouse gases and forcing. Since the emissions of CH_4, CO, NO_x, and VOCs, which strongly affect the formation of ozone, are maximized in the A2p scenario, the modelled forcings should represent an upper bound for the forcing produced under more constrained emissions scenarios. The eleven models simulate an increase in tropospheric ozone of 11.4 to 20.5 DU by 2100 corresponding to a range of radiative forcing from 0.40 to 0.78 W m^{-2}. Under this scenario, stratospheric ozone increases by between 7.5 to 9.3 DU, which raises the radiative forcing by an additional 0.15 to 0.17 W m^{-2}. The growth rates of atmospheric methane have fallen in the recent past, and recent projections of methane concentrations and radiative forcing for the period 2000 to 2030 are comparable to, or less than, the B2 SRES scenario (see Section 10.4.2).

One aspect of future direct aerosol radiative forcing omitted from all but 2 (the NASA GISS-EH and -ER models) of the 23 AOGCMS analyzed in IPCC AR4 is the role of nitrate aerosols. Rapid increases in emissions of NO_x could produce enough nitrate aerosol to offset the expected decline in sulphate forcing by 2100. Adams et al. (2001) have computed the radiative forcing by sulphate and nitrate accounting for the interactions among sulphate, nitrate, and ammonia. For 2000, the sulphate and nitrate forcing are –0.95 and –0.19 W m^{-2}, respectively. Under the SRES A2 scenario, by 2100 declining SO_2 emissions cause the sulphate forcing to drop to –0.85 W m^{-2}, while the nitrate forcing rises to –1.28 W m^{-2}. Hence the total sulphate-nitrate forcing increases from –1.14 W m^{-2} to –2.13 W m^{-2} rather than declining as models that omit nitrates would suggest.

Recent field programs focused on Asian aerosols have demonstrated the importance of BC and organic carbon (OC) for regional climate, including potentially significant perturbations to the surface energy budget and hydrological cycle (Ramanathan et al., 2001). The SRES scenarios include time series for chemical precursors of sulphate aerosols, but the SRES scenarios do not prescribe future concentrations of BC and OC. As a result, modelling groups have developed a multiplicity of projections for the concentrations of these aerosol species. For example, Takemura et al. (2001) use data sets for BC released by fossil fuel and biomass burning (Cooke and Wilson, 1996) under current conditions and scale them by the ratio of future to present-day CO_2. The emissions of OC are derived using OC:BC ratios estimated for each source and fuel type. Koch (2001) also employs scaling of present-day emissions inventories by the ratio of future to present-day CO_2 emissions. There are still very large uncertainties in current inventories of BC and OC (Bond et al., 2004), the ad hoc scaling methods used to produce future emissions, and the enormous disparity among various treatments of the optical properties of carbonaceous species. Given these uncertainties, future projections of forcing by BC and OC should be quite model dependent, even for a particular SRES emissions scenario.

The SRES scenarios did not explicitly consider changes in composition of the upper troposphere and lower stratosphere. Recent evidence suggests that there are detectable anthropogenic increases in stratospheric sulphate (e.g., Myhre et al., 2004), water vapor (e.g., Forster and Shine, 2002), and condensed water in the form of aircraft contrails. However, recent modelling studies suggest that these forcings are relatively minor compared to the major LLGHGs and aerosol species. Marquart et al. (2003) estimate that the radiative forcing by contrails will increase from 0.035 W m^{-2} in 1992, to 0.094 W m^{-2} in 2015, and to 0.148 W m^{-2} in 2050. The rise in forcing is due to an increase in subsonic aircraft traffic following estimates of future fuel consumption (Penner et al., 1999). These estimates are still subject to considerable uncertainties related to poor constraints on the microphysical properties, optical depths, and diurnal cycle of contrails (Myhre and Stordal, 2001; 2002; Marquart et al., 2003). Pitari et al. (2002) examine the effect of future emissions under the A2 scenario on stratospheric concentrations of sulphate aerosol and ozone. By 2030, the mass of stratospheric sulphate increases by approximately 33%, with the majority of the increase contributed by enhanced upward fluxes of anthropogenic SO_2 through the tropopause. The increase in direct shortwave forcing by stratospheric aerosols in the A2 scenario during 2000 to 2030 is –0.06 W m^{-2}.

Some recent studies have suggested that the global atmospheric burden of soil dust aerosols could decrease by between 20 and 60% due to reductions in desert areas associated with climate change (Mahowald and Luo, 2003). Tegen et al. (2004a; 2004b) compared simulations of ECHAM4 and HadCM3 including the effects of climate-induced changes in atmospheric conditions and vegetation cover and the effects of increased CO_2 concentrations on vegetation density. These simulations are forced with identical (IS92a) time series for well-mixed greenhouse gases. Their findings suggest that future projections of changes in dust loading are quite model dependent, since the net changes in global atmospheric dust loading produced by the two models have opposite signs. They also conclude that dust from agriculturally disturbed soils is less than 10% of the current burden, and that climate-induced changes in dust concentrations would dominate land-use changes under both minimum and maximum estimates of increased agricultural area by 2050.

10.3 Projected Changes in the Physical Climate System

The context for the climate change results presented here has been set in Chapter 8 (evaluation of simulation skill of the control runs and inherent natural variability of the global coupled climate models), and in Chapter 9 (evaluation of the simulations of 20th century climate using the global coupled climate models). A table describing the characteristics of the models was given in Chapter 8, and Table 10.3.1 summarizes the climate change experiments that have been performed with the AOGCMs and other models that will be assessed in this chapter.

The TAR showed multi-model results for future climate change from simple 1% per year CO_2 increase experiments, and from several scenarios including the older IS92a, and, new to the TAR, two SRES scenarios (A2 and B2). For the latter, results from nine models were shown for global averaged temperature change and regional changes. Since the TAR, an unprecedented internationally coordinated climate change experiment has been performed by 21 models from around the world as noted in Table 10.3.1. This larger number of models running the same experiments allows us to better quantify the multi-model signal as well as uncertainty regarding spread across the models (in this section), and also point the way to probabilistic estimates of future climate change (Section 10.5). The scenarios considered here include one of the SRES scenarios from the TAR, scenario A2, along with two additional scenarios, A1B and B1 (see Section 10.2 for details regarding the scenarios, and also Figure 10.3.1). This is a subset of the SRES marker scenarios used in the TAR, and they represent a "low" (B1), "medium" (A1B), and "high" (A2) scenario. This choice is made solely due to the limited computational resources for multi-model simulations using comprehensive AOGCMs and does not imply any preference or qualification of these three scenarios over the others.

Additionally, three climate change commitment experiments were performed, one where concentrations of GHGs were held fixed at year 2000 values and the models were run to 2100 (termed 20th century stabilization here), and two where concentrations were held fixed at year 2100 values for A1B and B1, and the models were run for an additional 100 to 200 years (see Section 10.7).

[INSERT FIGURE 10.3.1 HERE]

[INSERT TABLE 10.3.1 HERE]

This section considers the basic changes in climate over the next hundred years simulated by current climate models under plausible anthropogenic forcing scenarios. While we assess all studies in this field, the presentation will focus on results derived by the authors from the new data set for the three SRES scenarios considered in Section 10.2. Following TAR, we use means across the multi-model ensemble to illustrate representative changes. Studies such as Phillips and Gleckler (2006) have shown that such means are able to simulate the contemporary climate more accurately than individual models, and it is anticipated that this might be true for climate changes also. While we indicate the range of results here, the consideration of uncertainty resulting from this range is addressed more completely in Section 10.5. The use of means has the additional advantage of reducing the 'noise' associated with internal or unforced variability in the simulations. Models are equally weighted here, but other options are noted in Section 10.5.

Standard metrics for response of global coupled models are the equilibrium climate sensitivity, defined as the globally averaged surface air temperature change for a doubling of CO_2 for the atmosphere coupled to a non-dynamic slab ocean, and the transient climate response (TCR), defined as the globally averaged surface air temperature change at the time of CO_2 doubling in the 1% per year transient CO_2 increase experiment. The TAR showed results for these 1% simulations, and we discuss equilibrium climate sensitivity, TCR and other aspects of response in Section 10.5.2. Chapter 8 includes processes and feedbacks involved with these metrics.

The following subsections begin with the basic global warming signal relative to the contemporary climate period considered in Chapter 8. We then address patterns of change in warming, precipitation and quantities of particular relevance to impacts. Later subsections consider other important aspects of the climate system, with more reliance on published studies of the multi-model data set and similar results.

10.3.1 Time-Evolving Global Change

The globally averaged surface warming time series from each model in the multi-model data set is shown in Figure 10.3.2, either as a single member (if that was all that was available) or a multi-member ensemble mean, for each scenario in turn. The multi-model ensemble mean warming is also plotted for each case. The surface air temperature is used, averaged over each year, shown as an anomaly relative to the 1980–1999, and offset by any drift in the corresponding control runs in order to extract the forced response. The base period is chosen to match the contemporary climate simulation that is the focus of previous chapters. Similar results have been shown in studies of these models (e.g., Xu et al., 2005; Meehl et al., 2006b; Yukimoto et al., 2006). Interannual variability is evident for each single-model series, but little remains in the ensemble mean. This is because most of this is unforced and is a result of internal variability, as has been presented in detail in Section 9.2.2 of TAR. Clearly, there is a range of model results at each year, but over time this range becomes smaller relative to the mean warming. The range is somewhat smaller than the range of warming at 2100 for the A2 scenario in the comparable Figure 9.6 of TAR, despite the larger number of models here. Consistent with the range of forcing presented in 10.2, the warming by 2100 is largest in the high GHG growth scenario A2, intermediate in the moderate growth A1B, and lowest in the low growth B1. Naturally, models with high sensitivity tend to have above average warming in each scenario. Global mean precipitation increases in all scenarios (Figure 10.3.2, right column), indicating an intensification of the hydrological cycle. Douville et al. (2002) show that this is associated with increased water-holding capacity of the atmosphere, as the mean atmospheric residence time of the water vapour is actually increased. The multi-model mean varies approximately in proportion to the mean warming, though uncertainties in future hydrological cycle behaviour arise due in part to the different responses of tropical precipitation across models (Douville et al., 2005).

The trends of the multi-model mean temperature vary somewhat over the century because of the varying forcings, including that in aerosol (see Section 10.2). This is illustrated more clearly in Figure 10.3.2, which shows the mean warming series for each scenario as an extension of the 20th century simulations, assessed in Chapter 9 (Figure 9.4.1c). The time series beyond 2100 are derived from the extensions of the simulations (those available) under the idealised constant-forcing scenarios considered further in Section 10.7.1.

[INSERT FIGURE 10.3.2 HERE]

In order to focus on the forced response of the models at the regional scale, we reduce the internal variability further by averaging over 20-year time periods. This span is shorter than the traditional 30-year climatological period, in recognition of the transience of the simulations, and of the larger size of the ensemble. We focus on three periods over the coming century: an early century period 2011–2030, a mid-century period 2046–2065, and the late century period 2080–2099. Again, we consider changes of temperature, and other quantities that follow, relative to the 1980–1999 means. The multi-model ensemble mean warming for the three future periods in the different experiments are given in Table 10.3.2, among other results. The close agreement of warming for early century (with a range of only 0.05°C, from 0.64°C to 0.769°C) shows that no matter which scenario is followed, the warming is similar on the timescale of the next decade or two. Note that the precision given here is only relevant for comparison between these means. As evident in Figure 10.3.1, and discussed in 10.5, uncertainties in the projections are larger. It is also worth noting that nearly half of the early century climate change arises from warming we are already committed to (0.31°C for early century). By mid-century, the choice of scenario becomes more important for the magnitude of warming, with a range of 0.45°C from 1.28°C to 1.73°C, and with only about a quarter of that warming due to climate change we are already committed to (0.46°C). By the late century, there are clear consequences for which scenario is followed, with a range of 1.35°C from 1.77°C to 3.12°C, with only about 15% of that warming coming from climate change we are already committed to (0.55°C).

Table 10.3.2. Global mean warming (annual mean surface air temperature, in °C) for several time periods relative to 1980–1999 for the four scenarios simulated by the multi-model ensemble mean. Shown in italics are metrics related to the geographic patterns of warming (see Figure 10.3.5), first the M values for agreement of the normalized fields of warming, with the A1B 2080–2099 case, and second 100 times the mae (global mean absolute 'error'-difference, in °C) between the fields. Here $M = (200/\pi) \arcsin[1 - \mathrm{mse} / (V_X+V_Y + (G_X - G_Y)^2)]$, with mse the mean square error between the two fields X and Y, and V and G are variance and global mean of the fields (as subscripted). Values of 100 for M and 0 for mae indicate perfect agreement with the pattern of warming in A1B at the end of the 21st century.

	2011–2030	2046–2065	2080–2099	2180–2199
A2	0.64, *83, 8*	1.64, *91, 4*	3.12, *94, 3*	
A1B	0.69, *88, 5*	1.73, *94, 3*	2.62, *100, 0*	3.31, *89, 5*
B1	0.67, *86, 6*	1.28, *90, 4*	1.77, *93, 3*	2.11, *86, 6*
Commit	0.38, *76, 10*	0.46, *69, 12*	0.55, *71, 12*	

10.3.2 Patterns of Change in the 21st Century

10.3.2.1 Warming

It was noted in the TAR that much of the regional variation of the annual mean warming in the multi-model means is associated with high to low latitude contrast. We can better quantify this from the new multi-model mean in terms of zonal averages. A further contrast is provided by partitioning the land and ocean values based on model data interpolated to a standard grid. Figure 10.3.3 illustrates the late-century A2 case, with all values shown both in absolute terms, and also relative to the global mean warming. Warming over land is greater than the mean except in the southern midlatitudes, where the warming over ocean is a minimum. Warming over ocean is smaller than the mean except at high latitudes, where sea ice changes have an influence. This pattern of change illustrated by the ratios is quite similar across the scenarios. There is some contrast in the ratio with the commitment case (shown) to be considered in Section 10.7.1, which exhibits less contrast overall. At nearly all latitudes the A1B and B1 warming ratios lie between A2 and commitment, with A1B particularly close to the A2 results. Aside from the commitment case, the ratios for the other time periods are also quite similar to those for A2. We consider regional patterns and the precipitation contrasts shortly.

[INSERT FIGURE 10.3.3 HERE]

Zonal means also depict much of the atmospheric and oceanic variation of warming, and it is instructive to illustrate these features of the coupled system together. Figure 10.3.4 shows the warming for the A1B scenario at each latitude from the bottom of the ocean to the top of the atmosphere for the three 21st century periods used in Table 10.3.2. To produce this ensemble mean, the model data were first interpolated to standard ocean depths and atmospheric pressures. Consistent with the global transfer of excess heat from the atmosphere to the ocean, and the difference between warming over land and ocean, there is some discontinuity between the plotted means of the lower atmosphere and the upper ocean. The relatively uniform warming of the troposphere and cooling of the stratosphere in this multi-model mean is consistent with the changes shown in Chapter 9, Figure 9.8 of TAR, but now we also see its evolution during the 21st century. Upper tropospheric warming reaches a maximum in the tropics and is seen even in the early century time period. The pattern is very similar over the three periods, consistent with the rapid adjustment of the atmosphere to the forcing. These changes are simulated with good consistency among the models (the larger values of both signs are stippled, indicating that the ensemble mean is larger in magnitude than the inter-model standard deviation).

The ocean warming evolves more slowly. There is initially little warming below the mixed layer, except at some high latitudes. Even as a ratio with mean surface warming, later in the century the temperature increases more rapidly in the deep ocean, consistent with results from individual models (e.g., Watterson, 2003; Stouffer, 2004). This rapid warming of the atmosphere, and the slow penetration of the warming into the ocean has implications for the timescales of climate change commitment discussed in Section 10.7. It has been noted in a 5 member multi-model ensemble analysis that, associated with the changes in temperature of the upper ocean in Figure 10.3.4, the tropical Pacific ocean heat transport remains nearly constant with increasing GHGs due to the compensation of the subtropical cells (STCs) and the horizontal gyre variations, even as the STCs change in response to changes in the trade winds (Hazeleger, 2005). Additionally, a southward shift of the Antarctic Circumpolar Current is projected to occur in a 15 member multi-model ensemble due to changes of surface winds in a future warmer climate (Fyfe and Saenko, 2005). This is associated with a poleward shift of the westerlies at the surface (see Section 10.3.6), in the upper troposphere (particularly notable in the Southern Hemisphere, Stone and Fyfe, 2005), and increased relative angular momentum from stronger westerlies (Räisänen, 2003) and westerly momentum flux in the lower stratosphere particularly in the tropics and southern midlatitudes (Watanabe et al., 2005). The surface wind changes are associated with corresponding changes in wind stress curl and horizontal mass transport in the ocean (Saenko et al., 2005).

[INSERT FIGURE 10.3.4 HERE]

We turn to the regional warmings for each of the three scenarios and time period, shown as maps in Figure 10.3.5. In each case greater warming over most land is evident (e.g., Kunkel and Liang, 2005). Over the ocean warming is relatively large in the Arctic, and particularly pronounced along the equator in the Pacific (see Section 10.3.5.1) , with less warming to the north and south, and little warming over the North Atlantic and the Southern Ocean (e.g., Xu et al., 2005). Enhanced oceanic warming along the equator is evident in Figure 10.3.3, also. It can be associated with oceanic heat flux changes (Watterson, 2003) and forced by the atmosphere (Liu et al., 2005). It is clear that the pattern of change is very similar among the late century cases, with the pattern correlation coefficient as high as 0.994 between A2 and A1B. As for the zonal means, the fields normalized by the mean warming are very similar. The agreement between the A1B case, as a standard, and the others is quantified in Table 10.3.2, by the absolute measure M (Mielke, 1991; Watterson, 1996), with 100 meaning identical fields and zero meaning no similarity, being the expected value under random rearrangement of the data on the grid. Values of M become progressively larger later in the 21st century, with values of 90 or larger for the late 21st century, thus confirming the high similarity of the patterns of warming in the late century cases. The deviation from 100 is approximately proportional to the mean absolute difference also given. The earlier warming patterns are also similar to the standard case, particularly for the same scenario A1B. Furthermore, the zonal means over land and ocean considered above are representative of much of the small differences in warming ratio. While there is some influence of differences in forcing patterns among the scenarios, and of effects of oceanic uptake and heat transport in modifying the patterns over time, there is also support for the role of atmospheric heat transport in offsetting such influences (e.g., Boer and Yu, 2003b; Watterson and Dix, 2005). Dufresne et al. (2005) show that aerosol contributes a modest cooling of northern hemisphere up to the mid 21st century in the A2 scenario.

[INSERT FIGURE 10.3.5 HERE]

Such similarities in patterns of change have been described recently by Mitchell (2003) and Harvey (2004). They aid the efficient presentation of the broad scale multi-model results, as patterns depicted for the standard A1B 2080–2099 case are usually typical of other cases. To a large extent this applies to other seasons and also other variables under consideration here.. Where there is similarity of normalized changes, values for other cases can be estimated by scaling by the appropriate ratio of global means from Table 10.3.2.

The surface warming fields for the extratropical winter and summer seasons, December-February (DJF) and June-August (JJA), are shown for scenario A1B in Figure 10.3.6. The high latitude warming is rather seasonal, being larger in winter as a result of sea ice and snow as noted in Chapter 9 of the TAR. However, the relatively low warming in southern South America is more extensive in southern winter. Similar patterns of change in earlier model simulations are described by Giorgi et al. (2001).

[INSERT FIGURE 10.3.6 HERE]

The patterns of seasonal warming are similar to Figure 10.3.5 but they are enhanced in the winter hemisphere. They are consistent among models almost everywhere (almost the entire area of the globe is stippled).

10.3.2.2 Cloud and diurnal cycle

In addition to being an important link to humidity and precipitation, cloud cover plays an important role for the sensitivity of the GCMs (e.g., Soden and Held, 2006) and for the diurnal temperature range (DTR) over land (e.g., Dai and Trenberth, 2004 and references therein) so we consider the projection of these variables by multi-model ensembles. This was not shown in the TAR, and is made possible here by the new multi-model data set. Cloud radiative feedbacks to GHG forcing are sensitive to the elevation, latitude and hence temperature of the clouds, in addition to their optical depth and their atmospheric environment (see Chapter 8, Section 8.6.3.2). Current GCMs simulate clouds through various complex parameterizations (see Chapter 8, Section 8.2.1.3), to produce cloud cover quantified by an area fraction within each grid square, and each atmospheric layer. Taking zonal means of this quantity, averaged over the present and future periods produces a relative change that is indicative of the latitude-height structure of the cloud changes. Averaging across the multi-model ensemble (from available data), using results interpolated to standard pressure levels and latitudes, produces the changes depicted in Figure 10.3.7a. At all latitudes there are increases in the vicinity of the tropopause, and mostly decreases below, indicating an increase in the altitude of clouds overall. This shift occurs consistently across models. There are increases in near-surface amounts at some latitudes. There is considerable variation across the ensemble in these changes, as indicated by the small ratio of mean to standard deviation, with few areas stippled in the figure. The mid-level midlatitude decrease is rather consistent and also as much as a fifth of the average cloud fraction simulated for 1980-1999. It is worth noting that the stippled part in this figure and others under-represents mean changes that are formally statistically significant, if the individual model results were to be considered a sample.

The total cloud area fraction from an individual model represents the net coverage over all the layers, after allowance for the overlap of clouds, and is an output included in the data set. The change in the ensemble mean of this field is shown in Figure 10.3.7b. Much of the low and middle latitudes experience a small decrease in cloud cover. There are a few low latitude regions of increase, as well as substantial increases at high latitudes. The larger changes relate well to changes in precipitation depicted earlier. While clouds need not be precipitating, moderate spatial correlation between cloud cover and precipitation holds for seasonal means of both the present climate and changes.

[INSERT FIGURE 10.3.7 HERE]

The radiative effect of clouds is represented by the cloud radiative forcing diagnostic (see Chapter 8, Section 8.6.3.2). This can be evaluated from radiative fluxes at the top-of-atmosphere calculated with or without the presence of clouds, which are output by the GCMs. In the multi-model mean (not shown) values vary in sign over the globe. The global and annual mean averaged over the models, for 1980–1999, is –22.3 W m^{-2}.

Change in mean cloud radiative forcing has been shown to have different signs in a limited number of previous modelling studies (Meehl et al., 2004b; Tsushima et al., 2006). Figure 10.3.8a shows globally averaged cloud radiative forcing changes for the end of the A1B scenario for the last 20 years of the 21st century compared to the last 20 years of the 20th century for individual models of the data set. These current models show a variety of different magnitudes and even signs. The ensemble mean change is –0.6 W m^{-2}. This range indicates that cloud feedback is still an uncertain feature of the global coupled models (see Chapter 8, Section 8.6.3.2.2).

The diurnal range of surface air temperature (DTR) has been shown to be decreasing in several land areas of the globe in observations of the 20th century (see Chapter 3, Section 3.2.2.7), together with increasing cloud cover (see also Chapter 9, Section 9.4.2.3). In the multi-model mean of present climate DTR over land is indeed closely anti-correlated, spatially, to the total cloud cover field. This is true also of the 21st century changes in the fields, under A1B, as can be seen by comparing the change in DTR, shown as Figure 10.3.8b, with Figure 10.3.7b. Changes reach magnitude 0.5°C in some regions, with some consistency over the models. Smaller widespread decreases are likely due to the radiative effect of the enhanced greenhouse gases including water vapour (see also Stone and Weaver, 2002). Further consideration of DTR is given in Section 10.3.6.2.

In addition to the diurnal temperature range, Kitoh and Arakawa (2005) document changes in the regional patterns of diurnal precipitation over the Indonesian region, and show that over ocean nighttime precipitation decreases and daytime precipitation increases, while over land the opposite is the case, thus producing a decrease in the diurnal precipitation amplitude over land and ocean. They attribute these changes to a larger nighttime temperature increase over land due to increased GHGs.

[INSERT FIGURE 10.3.8 HERE]

10.3.2.3 Precipitation and surface water

Models simulate that global mean precipitation increases with global warming. However, there are substantial spatial and seasonal variations in this field even in the multi-model means depicted in Figure 10.3.6. There are fewer areas stippled for precipitation than for the warming, indicating more variation among the ensemble of models. Increases of precipitation at high latitudes in both seasons are very consistent across models. However, the increases of precipitation over the tropical oceans and in some of the monsoon regimes (e.g., South Asian monsoon in JJA, Australian monsoon in DJF) are notable but are not all consistent. There are smaller amplitude decreases of midlatitude summer precipitation. Decreases in precipitation over many subtropical areas are evident in the multi-model ensemble mean, but again are less consistent than the increases at high latitudes. Further discussion of regional changes is presented in Chapter 11.

With annual mean precipitation being of particular importance, the global map of the A1B 2080–2099 change is shown in Figure 10.3.9, along with some other hydrological quantities from the multi-model ensemble. Emori and Brown (2005) show percentage changes of annual precipitation from the ensemble. Increases of over 20% occur in most high latitudes, as well as eastern Africa, central Asia and the equatorial Pacific Ocean. The change over the ocean between 10°S and 10°N accounts for about half the increase in the global mean seen in Figure 10.3.1. Substantial decreases, reaching 20%, occur in the Mediterranean region, the Caribbean region, and the subtropical western coasts of each continent. Overall, changes over land account for 24% of the global mean increase in precipitation, a little less than the proportion of land by area (29%), but with local values of both signs.

These patterns of change occur in the other scenarios, although with agreement (by the metric M) a little lower than for the warming. The predominance of increases near the equator and at high latitudes, for both land and ocean, is clear from the zonal mean changes of precipitation included in Figure 10.3.3. The results for change scaled by global mean warming are rather similar across the four scenarios, an exception being a relatively large increase over the equatorial ocean for the commitment case. As with surface temperature, the A1B and B1 scaled values are always close to the A2 results. The zonal means of the percentage change map (shown in Figure 10.3.3) feature substantial decreases in the midlatitudes of both hemispheres in the A2 case, even if increases occur over some regions.

[INSERT FIGURE 10.3.9 HERE]

Wetherald and Manabe (2002) provide a good description of the mechanism of hydrological change simulated by GCMs. In GCMs the global mean evaporation changes closely balance the precipitation change, but not locally because of changes in the atmospheric transport of water vapour. Annual average evaporation (Figure 10.3.9) increases over much of the ocean, with spatial variations tending to relate to those in the surface warming (Figure 10.3.5). As found by Kutzbach et al. (2005) and Bosilovich et al. (2005), atmospheric moisture convergence increases over the equatorial oceans and over high latitudes. Over land, rainfall changes tend to be balanced by both evaporation and runoff. Runoff (Figure 10.3.9) is notably reduced in southern Europe and increased in south-east Asia and in high latitudes. The larger changes reach 20% or more of the simulated 1980–1999 values, which range from 1 to 5 mm d^{-1} in wetter regions to below 0.2 mm d^{-1} in deserts. (Note that runoff from the melting of ice sheets, Section 10.3.3, is not included here.) Nohara et al. (2006) and Milly et al. (2005) assess the impacts of these changes in terms of river flow, and find that discharges from high latitude rivers increase, while those from major rivers in the Middle East, Europe and central America tend to decrease.

While models simulate the moisture in the upper few meters of the land surface in varying ways, there is increasing confidence in multi-model means in representing large-scale changes (see Chapter 8, Section 8.3.4.2). In the annual mean, decreases in total soil moisture content (Figure 10.3.9) predominate, particularly in the subtropics. There are increases in some equatorial lands and northern Europe. Decreases also occur at high latitudes, where snow cover diminishes (Section 10.3.3). Regional hydrological changes are considered in Chapter 11 and also in the WGII report.

10.3.2.4 Sea-level pressure and atmospheric circulation

As a basic component of the mean atmospheric circulations and weather patterns, we consider projections of the mean sea-level pressure for the medium scenario A1B. Seasonal mean changes for DJF and JJA are shown in Figure 10.3.6 (matching results in Wang and Swail, 2006b). Sea level pressure differences show decreases at high latitudes in both seasons in both hemispheres, although the magnitudes of the changes vary (with no areas stippled). The compensating increases are predominantly over the midlatitude and subtropical ocean regions, extending across South America, Australia and southern Asia in JJA, and the Mediterranean in DJF. Many of these increases are consistent across the models. This pattern of change, discussed further in Section 10.3.5.3, has been linked to an expansion of the Hadley Circulation and a poleward shift of the midlatitude storm tracks (Yin, 2005). This helps explain, in part, the increases of precipitation at high latitudes and decreases in the subtropics and parts of the midlatitudes. Further analysis of the regional details of these changes is given in Chapter 11. The pattern of pressure change implies increased westerly flows across the western parts of the continents. These contribute to increases of mean precipitation (Figure 10.3.6) and increased precipitation intensity (Meehl et al., 2005a).

10.3.3 Changes in Ocean/Ice and High Latitude Climate

10.3.3.1 Changes in sea ice cover

Models of the 21st century project that future warming is amplified at high latitudes resulting from positive feedbacks involving snow and sea ice. The warming is particularly large in fall and early winter (Manabe and Stouffer, 1980; Holland and Bitz, 2003) when sea ice is thinnest and the snow depth is insufficient to blur the relationship between surface air temperature and sea ice thickness (Maykut and Untersteiner, 1971). As shown by Zhang and Walsh (2006), the coupled models show a range of responses in northern hemisphere sea ice areal extent ranging from very little change to a dramatic, and accelerating reduction over the 21st century (Figure 10.3.10a).

[INSERT FIGURE 10.3.10 HERE]

An important characteristic of the projected change is for summertime ice area to decline far more rapidly than wintertime ice area (Gordon and O'Farrell, 1997), and hence sea ice rapidly approaches a seasonal ice cover in both hemispheres (Figures 10.3.10b and 10.3.11). Seasonal ice cover is, however, rather robust and persists to some extent throughout the 21st century in most (if not all) models. Bitz and Roe (2004) noted that future projections show that Arctic sea ice thins fastest where it is initially thickest, a characteristic that future climate projections share with sea ice thinning observed in the late 20th century (Rothrock et al., 1999). Consistent with these results, a projection by Gregory et al. (2002b) showed that Arctic sea ice volume decreases more quickly than sea ice area (because trends in winter ice area are low) in the 21st century.

[INSERT FIGURE 10.3.11 HERE]

In 20th and 21st century simulations, Antarctic sea ice cover decreases more slowly than in the Arctic (Figure 10.3.11), particularly in the vicinity of the Ross Sea where most models predict a local minimum in surface warming. This is commensurate with the region with the greatest reduction in ocean heat loss, which results from reduced vertical mixing in the ocean (Gregory, 2000). The ocean stores much of its increased heat below 1 km depth in the Southern Ocean. In contrast, horizontal heat transport poleward of about 60°N increases in many models (Holland and Bitz, 2003), but much of this heat remains in the upper 1 km of the northern subpolar seas and Arctic Ocean (Gregory, 2000; Bitz et al., 2006). Bitz et al. (2006) argue that these differences in the depth where heat is accumulating in the high latitude oceans has consequences for the relative rates of sea ice decay in the Arctic and Antarctic.

While most climate models share these common characteristics (peak surface warming in fall and early winter, sea ice rapidly becomes seasonal, Arctic ice decays faster than Antarctic ice, and northward ocean heat transport increases into the northern high latitudes), models have poor agreement on the amount of thinning of sea ice (Flato and Participating CMIP modeling groups, 2004; Arzel et al., 2006) and the overall climate change in the polar regions (IPCC TAR) (Holland and Bitz, 2003). Flato (2004) showed that the basic state of the sea ice and the reduction in thickness and/or extent have little to do with sea ice model physics among CMIP2 models. At the same time, Holland and Bitz (2003) and Arzel et al. (2006) found serious biases in the basic state of simulated sea ice thickness and extent. Further, Rind et al. (1995), Holland and Bitz (2003), and Flato (2004) showed that the basic state of the sea ice had a significant influence on the change in sea ice thickness in the Arctic and extent in the Antarctic.

10.3.3.2 Other high latitude changes

Snow cover is an integrated response to both temperature and precipitation and exhibits strong negative correlation with air temperature in most areas with a seasonal snow cover (see Chapter 8, Sections 8.8.3.4 and 8.6.3.4 for an evaluation of model-simulated present day snow cover). Because of this temperature association, the simulations project widespread reductions in snow cover over the 21st century (Figure 10.3.12). For the Arctic Climate Impact Assessment (ACIA) model mean, at the end of the 21st century the projected reduction in the annual mean Northern Hemisphere snow coverage is –13% under the B2 scenario (ACIA, 2004). The individual model projections range from –9% to –17%. The actual reductions are greatest in spring and late autumn/early winter indicating a shortened snow cover season (ACIA, 2004). The beginning of the snow accumulating season (the end of the snow melting season) is projected to be later (earlier), and the fractional snow coverage is projected to decrease during the snow season (Hosaka et al., 2005).

[INSERT FIGURE 10.3.12 HERE]

Warming at high northern latitudes is also associated with large decreases in permafrost in climate model simulations (Lawrence and Slater, 2005; Yamaguchi et al., 2005; Kitabata et al., 2006). Lawrence and Slater (2005) estimate that over half of the area covered by the topmost layer of permafrost could thaw by 2050, and as much as 90% by 2100, with an increase of runoff to the Arctic Ocean of about 15% due to the melting of ground ice. Kitabata et al. (2006) analyze the same model and show comparable results, with a drying of summer soil moisture in late 21st century when significant amounts of permafrost have thawed. Similarly, Yamaguchi et al. (2005) show that as the upper layer permafrost begins to melt in a future wamer climate, soil moisture first increases during summer, but as more permafrost thaws the soil moisture decreases during summer. Stendel and Christensen (2002) show poleward movement of the extent of permafrost, and a 30–40% increase in active-layer thickness for most of the permafrost area in the Northern Hemisphere, with largest relative increases concentrated in the northernmost locations.

Regionally, the changes are a response to both increased temperature and increased precipitation (changes in circulation patterns) and are complicated by the competing effects of warming and increased snowfall in those regions that remain below freezing (see Chapter 4, Section 4.2 for a further discussion of processes that effect snow cover). Therefore, in contrast to the general Northern Hemisphere decrease of snow amount and snow coverage, some areas are projected to increase over parts of high-latitude northern regions during the cold seasons (Figure 10.3.12), and this is attributed to the increase of precipitation (snowfall) from autumn to winter (Hosaka et al., 2005). However, the projected snow coverage changes by the end of the 21st century are small (ACIA, 2004; Hosaka et al., 2005) and of comparable or smaller order than the present-day AR4 model bias (Hosaka et al., 2005; Roesch, 2006). We refer to Chapter 8, Section 8.3.4 for evaluation of present-day snow cover simulation by AR4 models and a further discussion of regional changes in Chapter 11, Section 11.3.8.

10.3.3.3 Changes in Greenland ice sheet mass balance

As noted in Section 10.6, modelling studies (e.g., Hanna et al., 2002; Kiilsholm et al., 2003; Wild et al., 2003) as well as satellite observations and airborne altimeter surveys (Krabill et al., 2000; Paterson and Reeh, 2001; Mote, 2003) suggest a slightly negative Greenland ice sheet mass balance associated with the thinning of ice sheet margins (see Chapter 5, Section 5.5.5.2). A consistent feature of all climate models is the projection of 21st century warming which is amplified in northern latitudes. This suggests a continuation of melting of the Greenland ice sheet, since increased summer melting dominates over increased winter precipitation in model projections of future climate. Ridley et al. (2005) coupled HadCM3 to an ice sheet model to explore the melting of the Greenland ice sheet under elevated (four times preindustrial) levels of atmospheric CO_2 (see Figure 10.6.4 below). While the entire Greenland Ice sheet eventually completely ablated (after 3000 years), the peak rate of melting was 0.1 Sv (see Sections 10.3.4 and 10.6.6). Toniazzo et al. (2004) further showed that in HadCM3, the complete melting of the Greenland Ice sheet was an irreversible process even if preindustrial levels of atmospheric CO_2 were re-established after its melting. Dethloff et al. (2004) went on to further show that Greenland's deglaciation had a profound influence on Arctic winter circulation with much less effect in the summer (consistent with Toniazzo et al., 2004).

10.3.4 Changes in the Meridional Overturning Circulation

A feature common to all climate model projections is the increase of high latitude temperature as well as an increase of high latitude precipitation. This was already reported in the IPCC TAR and is confirmed by the projections using the latest versions of comprehensive climate models (see Section 10.3.2). Both of these effects tend to make the high latitude surface waters lighter and hence increase their stability, thereby inhibiting convective processes. As more coupled models have become available since the TAR, the evolution of the Atlantic meridional overturning circulation (MOC) can be more thoroughly assessed. Figure 10.3.13 shows simulations from 15 coupled models integrated from 1850 to 2100 under SRES A1B atmospheric CO_2 and aerosol scenarios up to year 2100, and constant thereafter (see Figure 10.3.2). All of the models, except CCCma-CGCM3.1, INM-CM3.0 and MRI-CGCM2.3.2, were run without flux adjustments (see Chapter 8, Table 8.2.1). The MOC is influenced by the density structure of the Atlantic Ocean, small-scale mixing and the surface momentum and buoyancy fluxes. It is evident from Figure 10.3.13 that some models give a MOC strength that is inconsistent with the range of present-day estimates (Smethie and Fine, 2001; Ganachaud, 2003; Lumpkin and Speer, 2003; Talley, 2003). The MOC for these models is shown for completeness but cannot be used in assessing potential future changes in the MOC in response to various emissions scenarios.

[INSERT FIGURE 10.3.13 HERE]

Fewer studies have focused on projected changes in the Southern Ocean as a consequence of future climate warming. A common feature of coupled model simulations is the projected poleward shift and strengthening of the Southern Hemisphere westerlies (Yin, 2005; Fyfe and Saenko, 2006). This in turns leads to a strengthening, poleward shift and narrowing of the Antarctic Circumpolar Current. Fyfe and Saenko (Fyfe and Saenko, 2006) further noted that the enhanced equatorward surface Ekman transport, associated with the intensified westerlies, was balanced by an enhanced deep geostrophic poleward return flow below 2000 m.

Generally, the simulated late 20th century Atlantic MOC shows a spread ranging from a weak MOC of about 12 Sv (1 Sv = 10^6 m^3s^{-1}) to over 20 Sv (Figure 10.3.13, Schmittner et al., 2005). When forced with the SRES A1B scenario, the models show a reduction of the MOC, but in one model, the changes are not distinguishable from the simulated natural variability. The reduction of the MOC proceeds on the time scale of the simulated warming, because it is a direct response to the increase in buoyancy at the ocean surface. A positive NAO trend might delay, but not prevent, this response by a few decades (Delworth and Dixon, 2000). Such a weakening of the MOC in future climate causes reduced SST and salinity in the region of the Gulf Stream and North Atlantic Current (Dai et al., 2005). This can produce a decrease in northward heat transport south of 60°N, but increased northward heat transport north of 60°N (Hu et al., 2004a). No model shows an increase of the MOC in response to the increase in greenhouse gases, and no model simulates an abrupt reduction of the MOC within the 21st century. One study has suggested that inherent low frequency variability of the MOC, the Atlantic Multidecadal Oscillation, may produce a natural weakening over the next few decades that could further accentuate the decrease due to anthropogenic climate change (Knight et al., 2005, see Chapter 8, Section 8.4.6).

In some of the older models (e.g., Dixon et al., 1999), increased high latitude precipitation dominates over increased high latitude warming in causing the weakening, while in others (e.g., Mikolajewicz and Voss, 2000), the opposite is found. In a recent model intercomparison, Gregory et al. (2005) found that for all eleven models analysed, the MOC reduction was caused more by changes in surface heat flux than changes in surface freshwater flux. In addition, simulations using models of varying complexity (Stocker et al., 1992b; Saenko et al., 2003; Weaver et al., 2003) have shown that freshening or warming in the Southern Ocean acts to increase or stabilize the MOC. This is likely a consequence of the complex coupling of Southern Ocean Processes with North Atlantic Deep Water production.

A few simulations using coupled models are available which permit the assessment of the long-term stability of the MOC (Stouffer and Manabe, 1999; Voss and Mikolajewicz, 2001; Stouffer and Manabe, 2003; Wood et al., 2003; Yoshida et al., 2005; Bryan et al., 2006). Most of these simulations assume an idealized increase of CO_2 by 1%/year to various levels ranging from 2 to 4 times preindustrial levels. One study also considers slower increases (Stouffer and Manabe, 1999), or a reduction of CO_2 (Stouffer and Manabe, 2003). The more recent models are not flux adjusted and have high resolution (T85; 1.0°) (Yoshida et al., 2005; Bryan et al., 2006). A common feature of all simulations is a reduction of the MOC in response to the warming and a stabilization or recovery of the MOC when the concentration is kept constant after achieving a level of 2 to 4 times the preindustrial atmospheric CO_2 concentration. None of these models shows a spin-down of the MOC which continues after the forcing is kept constant. But such a long-term shut-down cannot be excluded if the amount of warming and its rate exceed certain thresholds as shown using a model of intermediate complexity (Stocker and Schmittner, 1997). Complete shut-downs, although not permanent, were also simulated by a flux adjusted coupled model (Manabe and Stouffer, 1994; Stouffer and Manabe, 2003; see also Chan and Motoi, 2005). Since in none of these simulations the thresholds determined by the model of intermediate complexity were passed (Stocker and Schmittner, 1997), the long-term stability of the MOC found in the present simulations is consistent with the results from simpler models.

The reduction in MOC strength associated with increasing greenhouse gases represents a negative feedback for the warming in and around the North Atlantic. That is, through reducing the transport of heat from low to high latitudes, SSTs are cooler than they would otherwise be if the MOC was left unchanged. As such, warming is reduced over and downstream of the North Atlantic. It is important to note that in models where the MOC weakens, warming still occurs downstream over Europe due to the overall dominant role of the radiative forcing associated with increasing greenhouse gases. Many future projections show that once the radiative forcing is held fixed, reestablishment of the MOC occurs to a state similar to that for the present day (Gregory et al., 2005). During this slow reestablishment phase, the MOC acts as a positive feedback to warming in and around the North Atlantic and, at equilibrium, there is close to zero net feedback. While the oceanic meridional heat flux at low latitude reduced upon a slowdown of the MOC, many simulations show increasing meridional heat flux into the Arctic which contributes to accelerated warming and sea ice melting there. This is due both to the advection of warmer water, as well as an intensification of the influx of North Atlantic water into the Arctic (Hu et al., 2004a).

Climate models for which a complete shutdown of the MOC has been found in response to sustained warming were flux adjusted coupled GCMs or intermediate complexity models. A robust result from such simulations is that the spin-down of the MOC takes several centuries after the forcing is kept fixed (e.g., at 4 $\times$ CO_2). Besides the forcing amplitude and rate (Stocker and Schmittner, 1997), the amount of mixing in the ocean also appears to determine the stability of the MOC: increased vertical and horizontal mixing tends to stabilize the MOC and to eliminate the possibility of a second equilibrium state (Manabe and Stouffer, 1999;

Longworth et al., 2005). Random internal variability or noise, often not present in simpler models, may also be important in determining the effective MOC stability (Monahan, 2002).

The MOC is not necessarily a comprehensive indicator of ocean circulation changes in response to global warming. In a transient 2 × CO_2 experiment using a coupled AOGCM, the MOC changes were small, but convection in the Labrador Sea stopped due to warmer, and hence lighter waters that inflow from the Greenland-Iceland-Norwegian Sea (GIN Sea) (Wood et al., 1999). Similar results were found by Hu et al. (2004a), who also report an increase in convection in the GIN Sea due to the influx of more saline waters from the North Atlantic. Various simulations using coupled models of different complexity find significant reductions in convection in the GIN Sea in response to warming (Schaeffer et al., 2004; Bryan et al., 2006). Presumably, a delicate balance exists in the GIN Sea between the circum-Arctic river runoff, sea ice production, and advection of saline waters from the North Atlantic, and on a longer time scale, the inflow of fresh water through Bering Strait. The projected increases in circum-Arctic river runoff (Wu et al., 2005) may enhance the tendency toward a reduction in GIN Sea convection (Stocker and Raible, 2005; Wu et al., 2005). Cessation of convection in the Labrador Sea in the next few decades is also simulated in a high-resolution model of the Atlantic Ocean driven by surface fluxes from two AOGCMs (Schweckendiek and Willebrand, 2005). The large-scale responses of the high-resolution ocean model (e.g., MOC, Labrador Seas) agree with those from the AOGCMs. The grid resolution of the ocean components in the coupled AOGCMs has significantly increased since the TAR, and some consistent patterns of changes in convection and water mass properties in the Atlantic Ocean emerge in response to the warming, but models still show a variety of responses in detail.

One of the most misunderstood issues concerning the future of the MOC under anthropogenic climate change is the popular notion that its reduction could cause the onset of an ice age. A relatively solid understanding of glacial inception exists wherein a change in seasonal incoming solar radiation (warmer winters and colder summers) associated with changes in the Earth's axial tilt, longitude of perihelion and the precession of its elliptical orbit around the sun is required (Crucifix and Loutre, 2002; Yoshimori et al., 2002). This small change must then be amplified by albedo feedbacks associated with enhanced snow/ice cover, vegetation feedbacks owing to the expansion of tundra, and greenhouse gas feedbacks associated with water vapour and the uptake (not release) of carbon dioxide and reduced release or increased destruction rate of methane. As discussed by Berger and Loutre (2002) and Weaver and Hillaire-Marcel (2004b), it is not possible for global warming to cause an ice age.

The best estimate of sea level from 1993–2003 (see Chapter 5, Section 5.5.5.2) associated with the slight net negative mass balance from Greenland is 0.1–0.2 mm/yr. This converts to only about 0.001–0.002 Sv of freshwater forcing over the total ocean. Such an amount, even when added directly and exclusively to the North Atlantic, has been suggested to be too small to affect the North Atlantic MOC (see Weaver and Hillaire-Marcel, 2004a). While one model exhibits a MOC weakening in the later part of the 21st century due to Greenland ice sheet melting (Fichefet et al., 2003), this same model had a very large downward drift of its overturning in the control climate, making it difficult to actually attribute the model MOC changes to the ice sheet melting. As noted in Section 10.3.3.3, Ridley et al. (2005) found the peak rate of Greenland Ice Sheet melting was 0.1 Sv when they instantaneously elevated Greenhouse gas levels in HadCM3. They further noted that this had little effect on the North Atlantic meridional overturning, although 0.1 Sv is sufficiently large to cause more dramatic transient changes in the strength of the MOC in other models (Stouffer et al., 2006).

Taken together, it is likely that the MOC will reduce, perhaps associated with a significant reduction in LSW formation, but very unlikely that the MOC will undergo an abrupt transition during the course of the 21st century. At this stage it is too early to assess the likelihood of an abrupt change of the MOC beyond the end of the 21st century. The few available simulations with models of different complexity rather suggest a centennial slow-down. Recovery of the MOC is likely if the radiative forcing is stabilised but would take several centuries.

10.3.5 Changes in Variability

10.3.5.1 Interannual variability in surface air temperature and precipitation

Future changes in anthropogenic forcing will result not only in changes in the mean climate state but also in the variability of climate. Addressing the interannual variability in monthly mean surface air temperature and precipitation of 19 AOGCMs in CMIP2, Räisänen (2002) found a decrease in temperature variability during the cold season in the extratropical Northern Hemisphere and a slight increase of temperature variability in low latitudes and in warm season northern mid latitudes. The former is likely due to the decrease of sea ice and snow with increasing temperature. The summertime decrease of soil moisture over the mid-latitude land surfaces contributes to the latter. Räisänen (2002) also found an increase in monthly mean precipitation in most areas, both in absolute value (standard deviation) and in relative value (coefficient of variation). Hu et al. (2000b) related this to increased variability of the tropical Pacific SST (El Niño variability) in their model, and Meehl and Arblaster (2003) linked changes of monsoon variability to non-linear increases of evaporation and consequent precipitation variability in the tropical Pacific. However, the significance level of these variability changes is markedly lower than that for time mean climate change. Similar results are obtained by 18 AOGCM simulations under the SRES A2 scenario (Giorgi and Bi, 2005).

10.3.5.2 Monsoons

In the tropics, an increase of precipitation is projected in the Asian monsoon and African monsoon in JJA, and the Australian monsoon in DJF in a warmer climate (Figure 10.3.6). The monsoonal precipitation in Mexico and Central America is projected to decrease in association with increasing precipitation over the eastern equatorial Pacific through Walker circulation and local Hadley circulation changes (Figure 10.3.6, also see Section 10.3.5.3). A more detailed assessment of regional monsoon changes is given in Chapter 11.

As global warming will lead to faster warming over land than over the oceans, the continental-scale land-sea thermal contrast will become larger in summer and become smaller in winter. Based on this, a simple idea is that the summer monsoon will be stronger and the winter monsoon will be weaker in the future than the present. However, model results are not as straightforward as this simple consideration. Tanaka et al. (2005) defined the intensities of Hadley, Walker and monsoon circulations using the velocity potential fields at 200 hPa. Using 15 AOGCMs, they showed a weakening of these tropical circulations by 9, 8 and 14%, respectively, by the late 21st century compared to the late 20th century. Using 8 AOGCMs, Ueda et al. (2006) demonstrated that pronounced warming over the tropics in the middle-to-upper troposphere causes a reduction in the meridional thermal gradient between the Asian continent and adjacent oceans, resulting in a weakening of monsoon circulations.

Despite weakening of the dynamical monsoon circulation, atmospheric moisture buildup due to increased GHGs and consequent temperature increase results in a larger moisture flux and more precipitation for the Indian monsoon (Douville et al., 2000; IPCC, 2001a; Ashrit et al., 2003; Meehl and Arblaster, 2003; May, 2004; Ashrit et al., 2005). For the South Asian summer monsoon, models suggest a northward shift of lower tropospheric monsoon wind systems with a weakening of the westerly flow over the northern Indian Ocean (Ashrit et al., 2003; 2005). Over Africa, multi-model analysis shows an increase in rainfall in the tropics and a decrease in the subtropics (see Section 10.3.2.3). The northward movement of the Sahara and the Sahel is expected due to the northward expansion and strengthening of the African monsoon (Liu et al., 2002; Haarsma et al., 2005).

Most model results project an increase of interannual variability in season-averaged Asian monsoon precipitation associated with an increase in its long-term mean value (e.g., Hu et al., 2000b; Räisänen, 2002; Meehl and Arblaster, 2003). Hu et al. (2000a) related this to increased variability of the tropical Pacific SST (El Niño variability) in their model. Meehl and Arblaster (2003) related the increased monsoon precipitation variability to increases of variability in evaporation and precipitation in the Pacific due to increased SSTs. Thus the South Asian monsoon variability is affected through the Walker circulation such that the role of the Pacific Ocean dominates and that of the Indian Ocean is secondary.

Loading of atmospheric aerosols affects regional climate and its future changes (see Chapter 7). If the direct effect of the aerosol increase is considered, surface temperatures will not get as warm because the aerosols reflect solar radiation. For this reason, land-sea temperature contrast becomes smaller than in the case without the direct aerosol effect, and the summer monsoon becomes weaker. Model simulations of the Asian monsoon show that the sulphate aerosols' direct effect reduces the magnitude of precipitation change compared with the case of only GHG increases (Emori et al., 1999; Roeckner et al., 1999; Lal and Singh, 2001). However, the relative cooling effect of sulfate aerosols is dominated by the effects of increasing GHGs by the end of the 21st century in the SRES marker scenarios (Figure 10.5.2). This results in the increased monsoon precipitation at the end of the 21st century in these scenarios (see Section 10.3.2.3). Furthermore, it is suggested that the aerosol with high absorptivity such as black carbon absorbs solar radiation in the lower atmosphere, cools the surface, stabilizes the atmosphere, and reduces precipitation (Ramanathan et al., 2001). The solar radiation reaching the surface decreases as much as 50% locally which could reduce the surface warming by GHGs (Ramanathan et al., 2005). These atmospheric brown clouds could make precipitation increase over the Indian Ocean in winter, decrease in the surrounding Indonesia region and the western Pacific Ocean (Chung et al., 2002), and reduce the summer monsoon precipitation both in South Asia and East Asia (Menon et al., 2002; Ramanathan et al., 2005). However, the total influence on monsoon preciptiation of time-varying direct and indirect effects of various aerosol species is still not resolved and the subject of active research.

10.3.5.3 ENSO changes in the tropics

Enhanced GHG concentrations result in a general increase in SST. These SST increases will not be spatially uniform, in association with general reduction in tropical circulations in a warmer climate (see Section 10.3.5.2). As shown in Figures 10.3.5 and 10.3.6, models have projected that the background tropical Pacific SST change from global warming (upon which individual ENSO events occur) will be an El Niño-like pattern. That is, the SST increase over the eastern tropical Pacific is larger than that over the western tropical Pacific, together with a decrease in SLP gradient along the equator and an eastward shift of the tropical Pacific rainfall distribution. Although individual models show a large scatter of "ENSO-ness" (Collins and The CMIP Modelling Groups, 2005; Yamaguchi and Noda, 2006), most realistic models show either no change in the mean state or a slight shift towards El Niño-like mean conditions (van Oldenborgh et al., 2005). Based on the spatial anomaly pattern of SST, SLP and precipitation, Yamaguchi and Noda (2006) showed that the CO_2-induced response pattern is closely related to the model natural variability, and in the tropical Pacific, an ENSO-like global warming pattern is simulated by many models with mostly an El Niño-like change (Figure 10.3.14). They suggested that the El Niño-like change may be attributable to the general reduction of tropical circulations due to the increased static stability in the tropics in a warmer climate (Knutson and Manabe, 1995; Sugi et al., 2002, Figure 10.3.4). In the models with the El Niño-like response, the positive feedback between SST, convection and atmospheric circulation (Bjerknes feedback) overwhelmed the negative cloud-radiation feedback (Jin et al., 2001; Yu and Boer, 2002). A shift towards mean El Niño-like conditions affects the whole tropics through Walker circulation changes. An eastward displacement of precipitation in the tropical Pacific accompanies an intensified and southwestward displaced subtropical anticyclone in the western Pacific, which can be effective to transport moisture from the low latitudes to the Meiyu/Baiu-region to bring more precipitation in East Asian summer monsoon (Kitoh and Uchiyama, 2006).

[INSERT FIGURE 10.3.14 HERE]

The projected change of the amplitude, frequency, and spatial pattern of El Niño itself is addressed next. The mean state change, through change in the sensitivity of SST variability to surface wind stress, plays a key role in determining the ENSO variance characteristics (Hu et al., 2004b; Zelle et al., 2005). For example, a more stable ENSO system is less sensitive to changes in the background state than one that is closer to instability (Zelle et al., 2005). Thus GCMs with an improper simulation of present-day climate mean state and air-sea coupling strength are not suitable for ENSO amplitude projections. Van Oldenborgh et al. (2005) categorized 19 models with their skill in the present-day ENSO simulations. Using the most realistic six out of 19 models, they find no statistically significant changes in amplitude of ENSO variability in the future. Large uncertainties in the skewness of the variability limits the assessment of the future relative strength of El Niño and La Niña events. Even with the larger warming scenario under $4 \times CO_2$ climate, Yeh and Kirtman (2005) find that despite the large changes in the tropical Pacific mean state, the changes in ENSO amplitude are highly model dependent. Meehl et al. (2006a) analysed two AOGCMs with increased GHGs and found that the decrease of El Niño amplitude in those two models was related to warming below the thermocline that weakened the stratification.

On the other hand, Guilyardi (2006) assessed mean state, coupling strength and modes (SST mode resulting from local SST-winds interaction or thermocline mode resulting from remote winds-thermocline feedbacks), using the pre-industrial control, stabilized 2 × CO_2 and 4 × CO_2 simulations in a multi-model ensemble. The models that exhibit the largest El Niño amplitude change in scenario experiments are those that shift towards a thermocline mode. The observed 1976 climate shift in the tropical Pacific actually involved such a mode shift (Fedorov and Philander, 2001). Most of those models that best simulate the tropical Pacific climatology in terms of mean state, seasonal cycle and coupling strength show the above mode change, implying an increasing likelihood of increased El Niño amplitude in a warmer climate. Merryfield (2006) also analysed a multi-model ensemble and found a wide range of behaviour for future El Niño amplitude, ranging from little change to larger El Niño events to smaller El Niño events, though several models that simulated some observed aspects of present-day El Niño events showed future increases in El Niño amplitude. However, significant multi-decadal fluctuations of El Niño amplitude in observations and long coupled model control runs add another complicating factor to attempting to discern whether any future changes of El Niño amplitude are due to external forcing or are simply a manifestation of internal multi-decadal variability (Meehl et al., 2006a). Therefore, there are no clear indications at this time regarding future changes of El Niño amplitude in a warmer climate.

For the change of ENSO frequency, Saenko (2006) showed a decrease in the time scale for large-scale dynamic oceanic adjustment based on increases in the first baroclinic Rossby radius of deformation due to oceanic stratification in warmer climate. These oceanic stratification changes and calculated increasing oceanic internal wave speeds are seen in most of the AOGCMs. However, multi-model analyses by Guilyardi (2006) find no clear indication of ENSO frequency change in a warmer climate.

The tropospheric biennial oscillation (TBO) has been suggested as a fundamental set of coupled interactions in the Indo-Pacific region that encompass ENSO and the Asian-Australian monsoon, and the TBO has been shown to be simulated in current AOGCMs (Chapter 8). Nanjundiah et al. (2005) analyse a multi-model dataset to show that, for models that successfully simulate the TBO for present-day climate, the TBO becomes more prominent in a future warmer climate due to changes in the base state climate, though, as with ENSO, there is considerable inherent decadal variability regarding the relative dominance of TBO and ENSO with time.

In summary, the mean tropical Pacific state tends to shift towards mean El Niño-like SST conditions with the eastern Pacific warming more than the western Pacific in a future warmer climate. A shift towards mean El Niño-like conditions accompanies a weakened Walker circulation and eastward displacement of precipitation in the tropical Pacific. There is a wide range of behaviour among the current models with no clear indication regarding possible changes of future El Niño amplitude or frequency.

10.3.5.4 ENSO-monsoon relationship

ENSO affects interannual variability in the whole tropics through changes in the Walker circulation. It has been known that there is a significant correlation between ENSO and tropical circulation/precipitation from the analysis of observational data. There is a tendency for less Indian summer monsoon rainfall in El Niño years, and above normal rainfall in La Niña years. Recent analyses have revealed that the correlation between ENSO and the Indian summer monsoon has decreased recently, and many hypotheses have been raised (see Chapter 3). With respect to global warming, one hypothesis is that the Walker circulation (accompanying ENSO) shifted south-eastward, reducing downward motion in the Indian monsoon region, which originally suppressed precipitation in that region at the time of El Niño, but now produces normal precipitation as a result (Krishna Kumar et al., 1999). Another explanation is that as the ground temperature of the Eurasian continent has risen in the winter-spring season due to global warming, the temperature difference between the continent and the ocean has become large, thereby causing more precipitation, and the Indian monsoon is normal in spite of the occurrence of El Niño (Ashrit et al., 2001).

It is reported that the MPI model (Ashrit et al., 2001) and the ARPEGE-OPA model (Ashrit et al., 2003) showed no global warming-related change in the ENSO-monsoon relationship, although a decadal-scale fluctuation is seen, suggesting a weakening of the relationship might be part of the natural variability. However, Ashrit et al. (2001) showed that while the impact of La Niña does not change, the influence of El Niño on the monsoon becomes small, suggesting the possibility of asymmetric behavior of the changes in the ENSO-monsoon relationship. On the other hand, the MRI-CGCM2 indicates a weakening of the correlation into the 21st century particularly after 2050 (Ashrit et al., 2005). The MRI-CGCM2 model results support the above hypothesis that the Walker circulation shifts eastward and no longer influences India at the time of El

Niño in a warmer climate. This eastward shift is the expected response of an El Niño-like climate change (see Section 10.3.5.3). Camberlin et al. (2004) extended their analysis to other ENSO-affected regions and found decadal fluctuations in ENSO's effect on regional precipitation. In most cases, these fluctuations may reflect natural variability of the ENSO teleconnection, and long-term correlation trends may be comparatively weaker. One possible reason is that the ARPEGE-OPA model simulated very regular ENSO events through the 21st century with little change in either the mean annual cycle or the monthly standard deviation.

In summary, the ENSO-monsoon relationship can vary from decade to decade purely due to internal variability of the climate system. In spite of the issues related to natural variability, there is some evidence of a future weakening of the ENSO-monsoon relationship in a future warmer climate.

10.3.5.5 Annular modes and mid-latitude circulation changes

Since the TAR, there have been a number of modeling studies that have investigated responses of extratropical climate variability to various anthropogenic and natural forcings with more comprehensive experiments including a larger size of ensemble simulations. Additionally, the analyses have incorporated model intercomparisons and model ensembles to reduce the uncertainties.

Many simulations project some decrease of the Arctic surface pressure in the 21st century, as seen in the multi-model average (see Figure 10.3.6), which contributes to an increase of indices in the Northern Annular Mode (NAM) or the Arctic Oscillation (AO), as well as the North Atlantic Oscillation (NAO) that is closely related with NAM in the Atlantic sector (see Chapter 8). This is also consistent with increased "AO-ness" shown in Figure 10.3.14. From the recent multi-model analyses, more than half of the models exhibit a positive trend of the NAM (Rauthe et al., 2004; Miller et al., 2006) and/or NAO (Osborn, 2004; Kuzmina et al., 2005). Although the magnitude of the trends shows a large variation among different models, Miller et al. (2006) found that none of the 14 models exhibits a trend toward a lower NAM index and higher Arctic SLP. In another multi-model analysis Stephenson et al. (2006) showed that of the 15 models able to simulate the NAO pressure dipole, 13 predicted a positive increase in the NAO with increasing CO2 concentrations, though the magnitude of the response was generally small and model-dependent. Only one study (McHugh and Rogers, 2005) suggests a weakened NAO circulation from increased GHGs from a 10 member multi-model ensemble. However, the multi-model average from the larger number (21) of models shown in Figure 10.3.6 indicates that it is likely that the NAM would not notably decrease in a future warmer climate. The average of IPCC-AR4 simulations from thirteen models suggests the increase becomes statistically significant early in the 21st century (Figure 10.3.15a, Miller et al., 2006).

[INSERT FIGURE 10.3.15 HERE]

Generally, models simulate a much smaller trend of NAM (or NAO) than that observed in the last half of 20th century (Gillett et al., 2002; Osborn, 2004), however the observed trend may contain considerable internal multi-decadal variability as implied from a notable decline of the index after the 1990s (see Chapter 3). Yukimoto and Kodera (2005) simulated a positive AO-like signal with ensemble AOGCM simulations for the 20th century, but the signal in response to GHG increase is about one third of the magnitude of decadal internal variability. Selten et al. (2004) also suggest that the observed NAO trend is an expression of a random climate variation, based on their 62 ensemble AOGCM simulations. It is suggested that such a decadal internal variability of NAO often involves a linkage with the tropical SST variations (e.g., Raible et al., 2005).

The spatial patterns of the simulated SLP trends vary among different models, in spite of close correlations of the models' leading patterns of inter-annual (or internal) variability with the observations (Osborn, 2004; Miller et al., 2006). However at the hemispheric scale of SLP change, the lowering in the Arctic is seen in the multi-model mean (Figure 10.3.6), though the change is smaller than the inter-model standard deviation. Besides the decrease in the Arctic region, increases over the Mediterranean Sea and the North Pacific exceed the inter-model standard deviation, the former suggests an association with northeastward shift of the NAO's center of action (Hu and Wu, 2004). The diversity of the patterns seems to reflect different responses in the Aleutian Low (Rauthe et al., 2004) in the North Pacific. Yamaguchi and Noda (2006) discussed the model response of ENSO versus AO, and find that many models project a positive AO-like change (Figure 10.3.14). In the North Pacific in high latitudes, however, the SLP anomalies are incompatible between the El

Niño-like change and the positive AO-like change, because models that project an El Niño-like change over the Pacific give a non-AO-like pattern in the polar region. As a result, the present models cannot fully determine the relative importance between the mechanisms inducing the positive AO-like change and inducing the ENSO-like change, leading to scatter in global warming patterns in regional scales over the North Pacific. Rauthe et al. (2004) suggest that the effects of sulfate aerosols contribute to a deepening of the Aleutian Low resulting in a slower or smaller increase of the AO.

Analyses of results from various models indicate that NAM can respond to increasing GHG concentrations through tropospheric processes (Fyfe et al., 1999; Gillett et al., 2003; Miller et al., 2006). Greenhouse gases can also drive a positive NAM trend through changes to the stratospheric circulation, similar to the mechanism by which volcanic aerosols in the stratosphere force positive annular changes (Shindell et al., 2001). Models with their upper boundaries extending farther into the stratosphere exhibit, on average, a relatively larger increase of the NAM and respond consistently to the volcanic forcing as observed (Figure 10.3.15a, Miller et al., 2006), implying the importance of the connection between the troposphere and the stratosphere.

A plausible explanation for the cause of the upward NAM trend in the models is an intensification of the polar vortex resulting from both tropospheric warming and stratospheric cooling mainly due to the increase of GHGs (Shindell et al., 2001; Sigmond et al., 2004; Rind et al., 2005a). The response may not be linear with the magnitude of radiative forcing (Gillett et al., 2002) since the polar vortex response is attributable to an equatorward refraction of planetary waves (Eichelberger and Holton, 2002) rather than radiative forcing itself. Since the long-term variation of the NAO is closely related with SST variations (Rodwell et al., 1999), it is considered to be essential that the projection of the changes in the tropical SST (Hoerling et al., 2004; Hurrell et al., 2004) and/or meridional gradient of the SST change (Rind et al., 2005b) should also be reliable. Rind et al. (2005b) suggested that it is likely that the current tendency for an increased positive phase of the AO/NAO will continue if there is significant tropical and high latitude warming.

The future trend of the Southern Annular Mode (SAM) or the Antarctic Oscillation (AAO) has been projected in a number of model simulations (Gillett and Thompson, 2003; Shindell and Schmidt, 2004; Arblaster and Meehl, 2006; Miller et al., 2006). According to the latest multi-model analysis (Miller et al., 2006), most models indicate a positive trend in the SAM index, and a lowering trend in the Antarctic SLP (as seen in Figure 10.3.6), with a higher likelihood than for the future NAM trend. On average, a larger positive trend is projected during the late twentieth century by models that include stratospheric ozone changes than those that do not (Figure 10.3.15b), though during the twenty-first century, when ozone changes are smaller, the SAM trends of models with and without ozone are similar. The cause of the positive SAM trend in the second half of the 20th century is mainly attributed to the stratospheric ozone depletion, evidenced by the fact that the signal is largest in the lower stratosphere in austral spring through summer (Thompson and Solomon, 2002; Arblaster and Meehl, 2006). However, increases of GHGs are also important factors (Shindell and Schmidt, 2004; Arblaster and Meehl, 2006) for the year-round positive SAM trend by accounting for the trend during early winter (Miller et al., 2006). During the twenty-first century, although the ozone amount is expected to stabilize or recover, the polar vortex intensification is likely to continue due to the increases of GHGs (Arblaster and Meehl, 2006).

It is implied that the future change of the annular modes leads to modifications of the future change in various fields such as surface temperatures, precipitation, and sea ice with regional features similar to those for the modes of natural variability (e.g., Hurrell et al., 2003). For instance, the surface warming in winter would be intensified in northern Eurasia and most of North America while weakened in the western North Atlantic, and the winter precipitation would increase in northern Europe while decreasing in southern Europe. The atmospheric circulation change would also affect the ocean circulations. Sakamoto et al. (2005) simulated an intensification of the Kuroshio but no shift of the Kuroshio extension, in response to an AO-like circulation change for the 21st century. However, Sato et al. (2006) simulated a northward shift of the Kuroshio extension, which leads to a strong warming off the eastern coast of Japan.

In summary, the future changes in the extratropical circulation variability are likely to be characterized by increases of positive phases in both the NAM and SAM. The response in the NAM to the anthropogenic forcing might not be distinct from the larger multi-decadal internal variability in the first half of the 21st century. The change in the SAM would appear earlier and more remarkably than the NAM since the stratospheric ozone depletion acts as an additional forcing. The positive trends of annular modes would influence the regional changes in temperature, precipitation and other various fields, similar to those accompanied by the NAM and SAM in the present climate, but would be superimposed on the global scale changes in a future warmer climate.

10.3.6 Future Changes in Weather and Climate Extremes

Projections of future changes of extremes are relying on an increasingly sophisticated set of models and statistical techniques. Studies assessed in this section rely on multi-member ensembles (3 to 5 members) from single models, analyses of multi-model ensembles ranging from 8 to 15 or more AOGCMs, and a perturbed physics ensemble with a single mixed layer model with over 50 members. The discussion here is intended to identify general characteristics of changes of extremes in a global context. Chapter 11 will address changes of extremes for specific regions.

10.3.6.1 Precipitation extremes

A long-standing result from global coupled models noted in the TAR was an increased chance of summer drying in the midlatitudes in a future warmer climate with associated increased risk of drought. This was noted in Figure 10.3.9, and has been documented in the more recent generation of models (Burke et al., 2006; Meehl et al., 2006b; Rowell and Jones, 2006). For example, Wang (2005) analyzed 15 recent AOGCMs to show that in a future warmer climate, the models simulate summer dryness in most parts of northern subtropics and midlatitudes, but there is a large range in the amplitude of summer dryness across models. Droughts associated with this summer drying could result in regional vegetation die-offs (Breshears et al., 2005) and contribute to an increase in the percentage of land area experiencing drought at any one time, for example, extreme drought increasing from 1% of present day land area (by definition) to 30% by the end of the century in the A2 scenario (Burke et al., 2006). Drier soil conditions can also contribute to more severe heat waves as discussed below (Brabson et al., 2005).

Associated with the risk of drying is also an increased chance of intense precipitation and flooding. Though somewhat counter-intuitive, this is because precipitation is concentrated into more intense events, with longer periods of little precipitation in between. Therefore, intense and heavy episodic rainfall events are interspersed with longer relatively dry periods with increased evapotranspiration, particularly in the subtropics as discussed further below in relation to Figure 10.3.17 (Frei et al., 1998; Allen and Ingram, 2002; Palmer and Räisänen, 2002; Christensen and Christensen, 2003; Beniston, 2004; Christensen and Christensen, 2004; Pal et al., 2004; Meehl et al., 2005a). However, increases in the frequency of dry days does not necessarily mean a decrease in the frequency of extreme high rainfall events depending on the threshold used to define such events (Barnett et al., 2006). Another aspect of these changes has been related to the mean changes of precipitation, with wet extremes becoming more severe in many areas where mean precipitation increases, and dry extremes where the mean precipitation decreases (Kharin and Zwiers, 2005; Meehl et al., 2005a; Räisänen, 2005a; Barnett et al., 2006). However, analysis of the 53 member perturbed physics ensemble indicates that the change in the frequency of extreme precipitation at an individual location can be difficult to estimate definitively due to model parameterization uncertainty (Barnett et al., 2006). Some specific regional aspects of these changes in precipitation extremes are discussed further in Chapter 11.

Climate models continue to confirm the earlier results that in a future climate warmed by increasing GHGs, precipitation intensity increases over most regions (Wilby and Wigley, 2002; Kharin and Zwiers, 2005; Meehl et al., 2005a; Barnett et al., 2006), and the increase of precipitation extremes is greater than changes in mean precipitation (Kharin and Zwiers, 2005). As discussed in Chapter 9, this is related to the fact that the energy budget of the atmosphere constrains increases of mean precipitation, but extreme precipitation relates to increases in moisture content and thus the non-linearities involved with the Clausius-Clapeyron relationship such that, for a given increase in temperature, increases in extreme precipitation can be more than the mean precipitation increase (e.g., Allen and Ingram, 2002). Additionally, increases in the frequency of seasonal mean rainfall extremes are greater than the increases in the frequency of daily extremes (Barnett et al., 2006). The increase of precipitation intensity in various regions has been attributed to contributions from both dynamic and thermodynamic processes associated with global warming (Emori and Brown, 2005), though in most regions the thermodynamic effect is dominant due to increases of water vapour, particularly in the extratropics. Meehl et al. (2005a) note the importance of the thermodynamic effect nearly everywhere, but points out that changes in circulation (dynamic effect) also make contributions to the pattern of precipitation intensity changes at mid and high latitudes. Recent studies have used more sophisticated statistical analyses involving extreme value theory that better quantify this result using a gamma distribution to study precipitation extremes. For example, Kharin and Zwiers (2005) showed that changes to both the location and scale of the extreme value distribution produced increases of precipitation extremes substantially greater than increases of annual mean precipitation. Watterson and Dix (2003) showed that an increase in the scale parameter from the gamma distribution represents an increase in precipitation intensity, and various regions such as the Northern Hemisphere land areas in winter showed particularly high values of increased scale parameter. This result was also seen in the study by Semenov and Bengtsson (2002). Time slice simulations with a higher resolution model (~1°) show similar results using changes in the gamma distribution, namely increased extremes of the hydrological cycle (Voss et al., 2002). However, there can also be some regional decreases, such as over the subtropical oceans (Semenov and Bengtsson, 2002).

Consistent with the results above for increased extremes of intense precipitation in many models, Watterson (2005) showed for the A2 scenario in the CSIRO model that though the storms in future climate in that model did not change much in intensity, there was an increase in extreme rainfall intensity with the extra-tropical surface lows, particularly over Northern Hemisphere land. This implies an increase of flooding. In a multi-model analysis of the CMIP models, Palmer and Räisänen (2002) showed that there was a three to five time increase in the likelihood of very wet winters over much of central and northern Europe due to an increase of intense precipitation associated with midlatitude storms suggesting increasing floods over Europe. They found similar results for summer precipitation with implications for more flooding in the Asian monsoon region in a future warmer climate. Similarly, Milly et al. (2002), Arora and Boer (2001) and Voss et al. (2002) related the increased risk of floods in a number of major river basins in a future warmer climate to an increase in river discharge. Christensen and Christensen (2003) concluded that there could be an increased risk of summertime flooding in Europe. McCabe et al. (2001) examined changes in Northern Hemisphere surface cyclones and came to similar conclusions regarding an enhanced risk for future intense storm-related precipitation events and flooding.

Global averaged time series of the Frich et al. (2002) indices in the multi-model analysis of Tebaldi et al. (2006) show simulated increases in precipitation intensity during the 20th century continuing through the 21st century (Figure 10.3.16), along with a somewhat weaker and less consistent trend for increasing dry periods between rainfall events for all scenarios (Figure 10.3.17). Part of the reason for these results is shown in the geographic maps for these quantities, where precipitation intensity increases almost everywhere, but particularly at mid and high latitudes where mean precipitation increases (Meehl et al., 2005a), (compare Figure 10.3.17 to Figure 10.3.6). However, in Figure 10.3.16 bottom, there are regions of increased runs of dry days between precipitation events in the subtropics and lower midlatitudes, but decreased runs of dry days at higher midlatitudes and high latitudes where mean precipitation increases (compare Figure 10.3.6 with Figure 10.3.16 bottom). Since there are areas of both increases and decreases of consecutive dry days between precipitation events in the multi-model average in Figure 10.3.6, the global mean trends are smaller and less consistent across models as shown in Figure 10.3.16. Consistency of response in a perturbed physics ensemble with one model shows only limited areas of increased frequency of wet days in July, and a larger range of changes of precipitation extremes relative to the ensemble mean in contrast to the more consistent response of temperature extremes (discussed below), indicating a less consistent response for preciptitation extremes in general compared to temperature extremes (Barnett et al., 2006).

[INSERT FIGURE 10.3.16 HERE]

[INSERT FIGURE 10.3.17 HERE]

10.3.6.2 Temperature extremes

The TAR concluded there was a very likely risk of increased temperature extremes, with more extreme heat episodes in a future climate. This result has been confirmed in subsequent studies (Yonetani and Gordon, 2001). Kharin and Zwiers (2005) show in a single model that future increases in temperature extremes follow increases in mean temperature over most of the world except where surface properties change (melting snow, drying soil). Furthermore, that study showed that in most instances warm extremes correspond to increases in daily maximum temperature, but cold extremes warm up faster than daily minimum temperatures, though this result is less consistent when model parameters are varied in a perturbed physics ensemble where there are increased daily temperature maxima for nearly the whole land surface. However, the range in magnitude of increases was substantial indicating a sensitivity to model formulations (Clark et al., 2006).

Weisheimer and Palmer (2005) examined changes in extreme seasonal (DJF and JJA) temperatures (exceeding the 95th percentile) in 14 models for 3 scenarios. They showed that by the end of 21st century, the probability of such extreme warm seasons rises above 90% in many tropical areas, and around 40% elsewhere. This result was confirmed in the perturbed physics ensemble where, for nearly all land areas, extreme JJA temperatures were at least 20 times and in some areas 100 times more frequent, making these changes many times greater than the ensemble spread and thus constituting a very robust result.

The TAR said nothing about possible changes in future cold air outbreaks. Vavrus et al. (2006) have analysed 7 AOGCMs run with the A1B scenario, and defined cold air outbreak as 2 or more consecutive days when the daily temperatures were at least 2 standard deviations below the winter-time mean. For a future warmer climate, they documented a decline in frequency of 50 to 100% in NH winter in most areas, with the smallest reductions occurring in western North America, the North Atlantic, and southern Europe and Asia due to atmospheric circulation changes associated with the increase of GHGs.

There were no studies at the time of the TAR that specifically documented changes in heat waves. Several recent studies have addressed possible future changes in heat waves explicitly, and found that in a future climate there is an increased risk of more intense, longer-lasting and more frequent heat waves (Meehl and Tebaldi, 2004; Schär et al., 2004; Clark et al., 2006). Meehl and Tebaldi (2004) showed that the pattern of future changes of heat waves, with greatest increases of intensity over western Europe and the Mediterranean, the southeast and western U.S., was related in part to base state circulation changes due to the increase in GHGs. An additional factor for extreme heat is drier soils in a future warmer climate (Brabson et al., 2005; Clark et al., 2006). Schär et al. (2004), Stott et al. (2004) and Beniston (2004) used the European 2003 heat wave as an example of the types of heat waves that are likely to become more common in a future warmer climate. Schär et al. (2004) and Weisheimer and Palmer (2005) noted that the increase in the frequency of extreme warm conditions was also associated with a change in interannual variability, such that the statistical distribution of mean summer temperatures is not merely shifted towards warmer conditions but also becomes wider. A multi-model ensemble shows that heat waves are simulated to have been increasing for the latter part of the 20th century, and are projected to increase globally and over most regions (Figure 10.3.17, Tebaldi et al., 2006), though different model parameters can contribute to the range in the magnitude of this response (Clark et al., 2006).

A decrease in diurnal temperature range in most regions in a future warmer climate was reported in the TAR. This has been re-confirmed by more recent studies (e.g., Stone and Weaver, 2002), and is discussed in Section 10.3.2.2 with more specific regional results assessed in Chapter 11. For a quantity related to the diurnal temperature range, it was concluded in the TAR that it would be likely that a future warmer climate would also be characterized by a decrease in frost days, though there were no studies at that time from global coupled climate models that addressed this issue explicitly. Since then it has been shown that there would indeed be decreases in frost days in a future climate in the extratropics (Meehl et al., 2004a), with the pattern of the decreases dictated by the changes in atmospheric circulation from the increase in GHGs (Meehl et al., 2004a). Results from an 8 member multi-model ensemble show simulated decreases in frost days for the 20th century continuing into the 21st century globally and in most regions (Figure 10.3.17). A quantity related to frost days is growing season length as defined by Frich et al. (2002), and this has been projected to increase in future climate (Tebaldi et al., 2006). This result is also shown in an 8 member multi-model ensemble where the simulated increase in growing season length in the 20th century continues into the 21st century globally and in most regions (Figure 10.3.17).

10.3.6.3 Tropical cyclones (hurricanes)

Earlier studies assessed in the TAR showed that future tropical cyclones would likely become more severe with greater wind speeds and more intense precipitation. More recent modelling experiments have addressed possible changes of tropical cyclones in a warmer climate and generally confirmed those earlier results. These studies fall into two categories: those with model grid spacings that only roughly represent some aspects of individual tropical cyclones, and those with model grid spacing of sufficient resolution that individual tropical cylones are reasonably simulated.

In the first category, there have been a number of climate change experiments with global models that can begin to simulate some characteristics of individual tropical cyclones, though studies with classes of models with 100 km resolution or higher cannot simulate observed tropical cyclone intensities due to the limitations of the relatively coarse grid spacing (e.g., Yoshimura and Sugi, 2006). A study with roughly 100 km grid spacing (T106) showed a decrease in tropical cyclone frequency globally but a regional increase over the North Atlantic and no significant changes in intensity (Sugi et al., 2002). Another study with that same resolution model indicated decreases in tropical cyclone frequency and intensity but more mean and extreme precipitation from the tropical cyclones simulated in the future in the western north Pacific (Hasegawa and Emori, 2005). Yoshimura and Sugi (2006) demonstrated in a sensitivity experiment with increased CO_2 in a T106 model that the decrease in tropical cyclone numbers was mainly due to the more stable tropical troposphere with greater warming in the upper troposphere compared to the surface, and the decrease in mean precipitation associated with reduction in the longwave cooling with the higher CO_2 levels. Yoshimura et al. (2006) further documented these decreases regionally and showed they occurred even with two different convection schemes in their model, with increased precipitation near the storm centers in the future. An AOGCM analysis with a more coarse resolution atmospheric model (T63, or about 200 km grid spacing) showed little change in overall numbers of the representations of tropical storms in that model, but a slight decrease in medium intensity storms in a warmer climate (Bengtsson et al., 2006). In another global modelling study with roughly a 100 km grid spacing, there was a 6% decrease in tropical storms globally and a slight increase in intensity, with both increases and decreases regionally related to the El Niño-like base state response in the tropical Pacific to increased GHGs (McDonald et al., 2005). In a global warming simulation with a T42 atmospheric model, the frequency of global tropical cyclone occurrence does not show significant changes (Tsutsui, 2002).

In the second category, studies have been performed with models that have been able to credibly simulate many aspects of tropical cyclones. Knutson and Tuleya (2004) used a high resolution (down to 9 km) mesoscale hurricane model with forcing from global coupled climate models with doubled CO_2 to show future tropical cyclones have 14% more intense central pressures, a 6% increase in maximum wind speeds, and an 18% increase in precipitation. That study also provided more details regarding how the choice of convective scheme can affect the quantitative nature of the results, though the main conclusions remain the same as in the earlier studies. Using a multiple nesting technique, an AOGCM was used to force a regional model over Australasia and the western Pacific with 125 km grid resolution, with an embedded 30 km resolution model over the southwestern Pacific (Walsh et al., 2004). At that 30 km resolution, the model is able to closely simulate the climatology of the observed tropical cyclone lower wind speed threshold of 17 m s^{-1}. Tropical cyclone occurrence (in terms of days of tropical cyclone activity) is slightly greater than observed, and the somewhat weaker than observed pressure gradients near the storm centers are associated with lower than observed maximum wind speeds, likely due to the 30 km grid spacing that is too coarse to capture extreme pressure gradients and winds. For $3 \times CO_2$ in that model configuration, the simulated tropical cylones experienced a 56% increase in maximum windspeed for winds greater than 30 m s^{-1}, and a 26% increase in the number of storms with central pressures less than 970 hPa, with no large changes in frequency and movement of tropical cyclones for that southwest Pacific region. It should also be noted that ENSO fluctuations have a strong impact on patterns of tropical cyclone occurrence in the southern Pacific (Nguyen and Walsh, 2001), and that uncertainty with respect future ENSO behaviour (Section 10.3.5.1) contributes to uncertainty with respect tropical cyclones (Walsh, 2004).

In another experiment, a global 20 km grid atmospheric model was run in time slice experiments for a present-day 10 year period and a 10 year period at the end of the 21st century for the A1B scenario to examine changes in tropical cyclones. Observed climatological SSTs were used to force the atmospheric model for the 10 year period at the end of the 20th century, and time-mean SST anomalies from an AOGCM simulation for the future climate were added to observed SSTs, and atmospheric composition was changed in the model to be consistent with the A1B scenario. At that resolution, tropical cyclone characteristics, numbers, and tracks were relatively well-simulated for present-day climate, though simulated intensities were somewhat weaker than observed (Oouchi et al., 2006). In that study, tropical cyclone frequency decreased 30% globally (but increased about 34% in the North Atlantic). The strongest tropical cyclones with extreme surface winds increased in number while weaker storms decreased. The tracks were not appreciably altered, and there was about a 14% increase in the maximum peak wind speeds in future simulated tropical cyclones in that model. As noted above, the competing effects of greater stabilization of the tropical troposphere (less storms) and greater SSTs (the storms that form are more intense) likely contribute to these changes except for the tropical North Atlantic where there are greater SST increases than in the other basins in that model. Therefore, the SST warming has a greater effect than the vertical stabilization in the Atlantic and produces not only more storms but more intense storms there. However, these regional changes are largely dependent on the spatial pattern of future simulated SST changes (Yoshimura et al., 2006).

Sugi et al. (2002) showed that the global-scale reduction in tropical cyclone frequency is closely related to weakening of tropospheric circulation in the tropics in terms of vertical mass flux. They noted that a significant increase in dry static stability in the tropical troposphere and little increase in tropical precipitation (or convective heating) are the main factors contributing to the weakening of the tropospheric circulation. Sugi and Yoshimura (2004) investigated a mechanism of this tropical precipitation change. They showed that the effect of CO_2 enhancement (without changing SST conditions) is a decrease in mean precipitation (Sugi and Yoshimura, 2004) and a significant decrease in the number of tropical cyclones as simulated in a T106 atmospheric model (Yoshimura and Sugi, 2006).

A synthesis of the model results to date indicates, for a future warmer climate, increased peak wind intensities and increased mean and peak precipitation intensities in future tropical cyclones, with the possibility of a decrease in the number of relatively weak hurricanes, and increased numbers of intense hurricanes and a global decrease in total numbers of tropical cyclones.

10.3.6.4 Extratropical storms and ocean wave height

It was noted in the TAR that there could be a future tendency for more intense extratropical storms, though the numbers could be less. This has been addressed by more recent studies. Some have shown little change in extratropical cyclone characteristics (Kharin and Zwiers, 2005; Watterson, 2005). But a tendency towards more intense systems was noted particularly in the A2 scenario in another global coupled climate model analysis (Leckebusch and Ulbrich, 2004), with more extreme wind events in association with those deepened cyclones for several regions of Western Europe, with similar changes in the B2 simulation though less pronounced in amplitude. Geng and Sugi (2003) used a higher resolution AGCM (T106) with time-slice experiments and obtained a decrease of cyclone density in both hemispheres in a warmer climate in both the DJF and JJA seasons, associated with the changes in the baroclinicity in the lower troposphere, in general agreement with coarser GCM results (e.g., Dai et al., 2001a), but density of strong cyclones increased while the density of weak and medium-strength cyclones decreased. Consistent with those results, several studies have shown a possible reduction of midlatitude storms in the Northern Hemisphere but an increase in intense storms (Lambert and Fyfe, 2006, for a 15 member multi-model ensemble) and for the Southern Hemisphere (Fyfe, 2003, with a possible 30% reduction in sub-Antarctic cyclones). More regional aspects of these changes were addressed for the Northern Hemisphere in a single model study by Inatsu and Kimoto (2005) who showed a more active storm track in the western Pacific in the future but weaker elsewhere. Fischer-Bruns et al.(2005) documented storm activity increasing over the North Atlantic and Southern Ocean, and decreases over the Pacific Ocean.

In agreement with earlier results (e.g., Schubert et al., 1998) a number of more recent studies have documented a poleward shift of several degrees longitude in midlatitude storm tracks (Geng and Sugi, 2003; Fischer-Bruns et al., 2005; Yin, 2005; Bengtsson et al., 2006). Consistent with these shifts in storm track activity, Cassano et al. (2006), using a 10 member multi-model ensemble, showed a future change to a more cyclonically-dominated circulation pattern in winter and summer over the Arctic, and increasing cyclonicity and stronger westerlies in the same multi-model ensemble for the Antarctic (Lynch et al., 2006).

By analyzing stratosphere-troposphere exchanges using time-slice experiments with the middle atmosphere version of ECHAM4, Land and Feichter (2003) suggested that cyclonic and blocking activity becomes weaker poleward of 30°N in a warmer climate at least in part due to decreased baroclinicity below 400 hPa, while cyclonic activity becomes stronger in the Southern Hemisphere associated with increased baroclinicity above 400 hPa. The atmospheric circulation variability on the interdecadal time scales may also change by increasing GHG and aerosols. One model result (Hu et al., 2001) showed that interdecadal variability of the SLP and 500 hPa height fields increased over the tropics and decreased in high latitudes by global warming.

The most consistent results from the current generation of models shows, for a future warmer climate, a poleward shift of storm tracks in both hemispheres, with generally fewer but more intense storms particularly in the Southern Hemisphere.

A new feature that has been studied related to extreme conditions over the oceans is wave height. Studies by Wang et al. (2004), Wang and Swail (2006a; 2006b), and Caires et al. (2006) have shown that for most regions of the midlatitude oceans, an increase of extreme wave height is likely to occur in a future warmer climate. This is related to increased wind speed associated with midlatitude storms, resulting in higher waves produced by these storms, and is consistent with the studies noted above that showed decreased numbers of midlatitude storms but more intense storms.

10.4 Carbon Cycle/Vegetation Feedbacks Ocean Acidification and Chemistry

10.4.1 Carbon Cycle/Vegetation Feedbacks

As a parallel activity to the standard IPCC AR4 climate projection simulations described in this chapter, the Coupled Climate Carbon Cycle Models Intercomparison Project (C4MIP) supported by WCRP and IGBP was initiated. Ten climate models with a representation of the land and ocean carbon cycle (see Chapter 7) performed simulations where the model is driven by anthropogenic CO_2 emissions scenario for the 1860–2100 time period (instead of atmospheric CO_2 concentration scenario as in the standard IPCC AR4 simulations). Each C4MIP model performed two simulations, a “coupled” simulation where the growth of atmospheric CO_2 induces a climate change which impacts on the carbon cycle, and an “uncoupled” simulation, where atmospheric CO_2 is radiatively inactive in order to estimate the atmospheric CO_2 growth rate one would get if the carbon cycle was unperturbed by the climate. Emissions were taken from the observation for the historical period (Houghton and Hackler, 2000; Marland et al., 2005) and from the IPCC SRES A2 scenario for the future (Leemans et al., 1998).

Chapter 7 describes the major results of the C4MIP models in terms of climate impact on the carbon cycle. Here we start from these impacts to infer the feedback on atmospheric CO_2 and therefore on the climate system. There is unanimous agreement amongst the models that future climate change will reduce the efficiency of the Earth system to absorb anthropogenic carbon dioxide essentially owing to a reduction of land carbon uptake. This latter is driven by a combination of reduced Net Primary Productivity and increased CO_2 soil respiration under a warmer climate. As a result, a growingly larger fraction of anthropogenic CO_2 will stay airborne if climate change controls the carbon cycle. By the end of the 21st century, this additional CO_2 varies between 20 ppm and 200 ppm for the two extreme models, most of the models lying between 50 and 100 ppm (Friedlingstein et al., 2006). This additional CO_2 leads to an additional radiative forcing ranging between 0.1 and 1.3 W m^{-2} and hence an additional warming ranging between 0.1 and 1.5°C.

In terms of atmospheric CO_2 evolution, the majority of the C4MIP models simulate a higher atmospheric CO_2 growth rate than the one taken as a forcing for the standard coupled models of the IPCC AR4 exercise (e.g., Meehl et al., 2005b). By 2100, atmospheric CO_2 varies between 730 and 1020 ppm for the C4MIP models, compared with 836 ppm for the SRES A2 concentration used by the standard IPCC-AR4 climate models (Figure 10.4.1a). The CO_2 concentration envelope of the C4MIP uncoupled simulations is centred on the standard SRES-A2 concentration value, the range reflecting the uncertainty in the carbon cycle. It should be noted that the standard SRES A2 concentration value of 836 ppm was calculated in the TAR with the BERN-CC model, accounting for the climate-carbon cycle feedback. Parameter sensitivity studies were performed with the BERN-CC model at that time and gave a range of 735 ppm to 1080 ppm, comparable to the range of the C4MIP study. The effects of climate feedback uncertainties on the carbon cycle have also been considered probabilistically by Wigley and Raper (2001). A later paper (Wigley, 2004) considered individual emissions scenarios, accounting for carbon cycle feedbacks in the same way as Wigley and Raper (2001). The results of these studies are consistent with the more recent C4MIP results. For the A2 scenario considered in C4MIP, the CO_2 concentration range in 2100 using the Wigley and Raper model is 769–1088 ppm, compared with 730–1020 ppm in the C4MIP study (which ignored the additional warming effect due to non-CO_2 gases). Similarly, using neural networks, Knutti et al. (2003) showed that the climate-carbon cycle feedback leads to an increase of 0.6°C of the central estimate for the SRES-A2 scenario.

[INSERT FIGURE 10.4.1 HERE]

Further uncertainties regarding carbon uptake were addressed in a 14 member multi-model ensemble using the CMIP2 models to quantify contributions to uncertainty with regards to inter-model variability versus internal variability (Berthelot et al., 2002). They found that the AOGCMs with the largest climate sensitivity also had the largest drying of soils in the tropics and thus the largest reduction of carbon uptake.

The C4MIP protocol did not account for the evolution of non-CO_2 greenhouse gases and aerosols. In order to compare the C4MIP simulated warming with the IPCC-AR4 climate models, we used the SRES A2 radiative forcings of CO_2 alone and total forcing (CO_2 plus non-CO_2 greenhouse gases and aerosols) as given in Appendix II of the TAR. Using these numbers and knowing the climate sensitivity of each C4MIP model, we can estimate the warming of the C4MIP models if they had included the non-CO_2 greenhouse gases and aerosols. Doing so, the C4MIP range of global temperature increase over the 21st century would be 2.4 to 5.6°C, compared with 2.6 to 4.1°C for standard IPCC-AR4 climate models (Figure 10.4.1b). As a result of a much larger CO_2 concentration by 2100 in the C4MIP models, the upper estimate of the global warming by 2100 is up to 1.5°C higher than for the SRES-A2 concentration standard simulations.

The C4MIP results highlight the importance of coupling the climate system and the carbon cycle in order to simulate, for a given scenario of CO_2 emission, a climate change that takes into account the dynamic evolution of the Earth's capacity to absorb the CO_2 perturbation.

Conversely, the climate-carbon cycle feedback will have an impact on the estimate of the projected CO_2 emissions leading to a stabilisation of atmospheric CO_2 at a given level. The TAR showed the range of future emissions for the WRE stabilisation concentration scenarios, using different model parametrization (including the climate-carbon feedback, Joos et al., 2001; Kheshgi and Jain, 2003). However, the emission reduction due to this feedback was not quantified. Similarly to the C4MIP protocol, coupled and uncoupled simulations have been recently performed in order to specifically evaluate the impact of climate change on the future CO_2 emissions required to achieve stabilisation (Matthews, 2005; Jones et al., 2006). Figure 10.4.2 shows the emissions required to achieve CO_2 stabilisation for the SP550, SP 750 and SP 1000 concentration scenarios as simulated by four climate-carbon cycle models. As detailed above, the climate-carbon cycle feedback reduces the land and ocean uptakes of CO_2, leading to a reduction of the emissions compatible with a given atmospheric CO_2 stabilization pathway. The higher the stabilization scenario, the larger the climate change, the larger the impact on the carbon cycle, and hence the larger the emission reduction. The uncertainty on the strength of the climate-carbon cycle feedback highlighted in the C4MIP analysis is also visible here. For the SP1000 scenario, according to the models, the emission reductions are between 2 and 7 GtC/yr that is 15 to 40% of the emissions estimated without accounting for the climate-carbon cycle feedback.

[INSERT FIGURE 10.4.2 HERE]

The current uncertainty on processes driving the land and the ocean carbon uptake will translate into an uncertainty in the future emissions of CO_2 required to achieve stabilization. In Figure 10.4.3, the carbon cycle related uncertainty is addressed using the Bern2.5CC carbon cycle model of intermediate complexity (Joos et al., 2001; Plattner et al., 2001) and the series of S450 to SP1000 CO_2-stabilization scenarios. The range of emission uncertainty has been achieved using identical assumptions as made in IPCC TAR, varying ocean transport parameters and parameterizations describing the cycling of carbon through the terrestrial biosphere. Results are thus very closely comparable, and the small differences can be largely explained by the different CO_2 trajectories applied and the use of a dynamic ocean model here compared to IPCC TAR.

[INSERT FIGURE 10.4.3 HERE]

The model results confirm that for stabilization of atmospheric CO_2, the emissions need to be reduced well below the year 2000 values in all scenarios. This is true for the full range of simulations covering carbon cycle uncertainty, even including the upper bound which is based on rather extreme assumptions of terrestrial carbon cycle process.

Cumulated emissions for the period from 2000 to 2100 (to 2300), range between 596 GtC (933 GtC) for SP450 and 1236 GtC (3052 GtC) for SP1000. The emission uncertainty is found to vary between –26% to

+28% about the reference cases in year 2100 and by –26% to +34% in year 2300, increasing with time. The range of uncertainty thus depends on the magnitude of the CO_2 stabilization level and the induced climate change. The additional uncertainty in projected emissions due to uncertainty in climate sensitivity is illustrated by two additional simulations with 1.5 and 4.5°C. The resulting emissions for this range of climate sensitivities lie within the range covered by uncertainty in processes driving the carbon cycle

Both the standard IPCC-AR4 and the C4MIP models ignore the effect of land cover change. However, as described in Chapters 2 and 7 past and future changes in land cover may affect the climate through several processes. First, they may change surface characteristics such as albedo. Second, they may affect the latent vs. sensible heat ratio and therefore impact on surface temperature. Third, they may induce atmospheric CO_2 emissions, and fourth, they can affect the capacity of the land to take up atmospheric CO_2. So far, no comprehensive coupled OAGCM has addressed these four components all together. Using AGCMs, Defries et al. (2004) studied the impact of future land cover change on the climate, while Maynard and Royer (2004) did a similar experiment on Africa only. Defries et al., (2002) forced the Colorado State University GCM (Randall et al., 1996) with AMIP climatological sea surface temperatures and with either the present-day vegetation cover or a 2050 vegetation map adapted from a low growth scenario of the IMAGE-2 model (Leemans et al., 1998). The study found that in the tropics and subtropics, replacement of forests by grassland or cropland leads to a reduction of carbon assimilation, and therefore of latent heat flux. This latter reduction leads to a surface warming of up to 1.5°C in deforested tropical regions. Using the ARPEGE-Climat AGCM (Déqué et al., 1994) with a higher resolution over Africa, Maynard et al. (2002), performed two experiments, one simulation with 2 × CO_2 SSTs taken from a previous ARPEGE transient SRES B2 simulation and present-day vegetation, and one with the same SSTs but the vegetation taken from a SRES B2 simulation of the IMAGE-2 model (Leemans et al., 1998). Similarly to Defries et al., (2002), they found that future deforestation in tropical Africa leads to a redistribution of latent and sensible heat that leads to a warming of the surface. However, this warming is relatively small (0.4°C) and represents about 20% of the warming due to the atmospheric CO_2 doubling.

Two recent studies further investigated to relative roles of future changes in greenhouse gases versus future changes in land cover. Using a similar model design as Maynard and Royer (2004), Voldoire (2006) compared the climate change simulated under a 2050 SRESB2 greenhouse gases scenario to the one under a 2050 SRES B2 land cover change scenario. They show that the relative impact of vegetation change to GHG concentration increase is of the order of 10%, and can reach 30% over localized tropical regions. In a more comprehensive study, Feddema et al. (2005) applied the same methodology for the SRES A2 and B1 scenario over the 2000–2100 period. Similarly they found no significant effect at the global scale, but a potentially large effect at the regional scale, such as a warming of 2°C by 2100 over the Amazon for the A2 land cover change scenario, coming with a reduction of the diurnal temperature range. The general finding of these studies is that the climate change due to land cover changes may be important relative to greenhouse gases at the regional level, where intense land cover change occurs. Globally, the impact of greenhouse gases concentration dominates over the impact of land cover change

10.4.2 Ocean Acidification due to Increasing Atmospheric Carbon Dioxide

Increasing atmospheric CO_2 concentrations lowers oceanic pH and carbonate ion concentrations, thereby decreasing the level of saturation of calcium carbonate (Feely et al., 2004). Surface ocean pH today is already 0.1 unit lower than preindustrial values. By the end of the century, it will become another 0.3–0.4 units lower under the IS92a scenario, which translates to a 100-150% increase in $[H^+]$. Simultaneously, carbonate ion concentrations will decrease, making it more difficult for marine calcifying organisms to form biogenic calcium carbonate (Raven et al., 2005). Experimental evidence suggests that if these trends continue, key marine organisms – such as corals and some plankton – will have difficulty maintaining their external calcium carbonate skeletons (e.g., Gattuso et al., 1998; Kleypas et al., 1999; Riebesell et al., 2000; Zondervan et al., 2001; Langdon et al., 2003; Orr et al., 2005).

Ocean acidification will eventually lead to undersaturation and dissolution of calcium carbonate in parts of the surface ocean. While Southern Ocean surface waters are projected to first exhibit undersaturation with regard to $CaCO_3$, low latitude regions will be affected as well (Figure 10.4.4). Recently, Orr et al. (2005) reported that conditions detrimental to high-latitude ecosystems could develop within decades, not centuries as suggested previously (Figure 10.4.5). These changes will not only influence the global carbon cycle, but also threaten marine organisms that form their exoskeletons out of $CaCO_3$, which are essential components of the marine food web (Raven et al., 2005).

[INSERT FIGURE 10.4.4 HERE]

[INSERT FIGURE 10.4.5 HERE]

10.4.3 Simulations of Future Evolution of Methane, Ozone, and Oxidants

Simulations using coupled chemistry-climate models indicate that the trend in upper stratospheric ozone changes sign sometime between 2000 and 2005 due to the gradual reduction in halocarbons. While ozone concentrations in the upper stratosphere decreased at a rate of 400 ppbv (–6%) per decade during 1980–2000, they are projected to increase at 100 ppbv (1–2%) per decade for 2000–2020 (Austin and Butchart, 2003). On longer timescales, simulations are showing significant changes in ozone and methane relative to current concentrations. The changes are related to a variety of factors, including increased emissions of chemical precursors; changes in gas-phase and heterogeneous chemistry; altered climate conditions due to global warming; and greater transport and mixing across the tropopause. The impacts on methane and ozone from increased emissions are a direct effect of anthropogenic activity, while the impacts of different climate conditions and stratosphere-troposphere exchange represent indirect effects of these emissions (Grewe et al., 2001).

The projections for ozone based upon scenarios with high emissions (IS92a, Leggett et al., 1992) and SRES A2 (Nakicenovic and Swart, 2000) indicate that the concentrations of tropospheric ozone will increase throughout the 21st century, primarily as a result of these emissions. Simulations for the period 2015 through 2050 find increases in O_3 of 20 to 25% (Grewe et al., 2001; Hauglustaine and Brasseur, 2001), and simulations through 2100 indicate that O_3 below 250 mb may grow by 40 to 60% (Stevenson et al., 2000; Grenfell et al., 2003; Zeng and Pyle, 2003). The primary species contributing to the increase in tropospheric O_3 are anthropogenic emissions of NO_x, CH_4, CO, and compounds from fossil fuel combustion. The photochemical reactions that produce smog are accelerated by increases of 2.6× in the flux of NO_x, 2.5 × in the flux of CH_4, and 1.8 × in CO in the A2 scenario. Approximately 91% of the higher concentrations in O_3 are related to direct effects of these emissions, with the remainder of the increase are attributable to secondary effects of climate change (Zeng and Pyle, 2003). These emissions also lead to higher concentrations of oxidants including OH, leading to a reduction in the lifetime of tropospheric methane by 8% (Grewe et al., 2001).

Since the projected growth in emissions occurs primarily in low latitudes, the ozone increases are largest in the tropics and sub-tropics (Grenfell et al., 2003). In particular, the concentrations in SE Asia, India, and Central America increase by 60 to 80% by 2050 under the A2 scenario. However, the effects of tropical emissions are not highly localized, since the ozone spreads throughout the lower atmosphere in plumes emanating from these regions. As a result, the ozone in remote marine regions in the southern hemisphere grows by 10 to 20% over present-day levels by 2050. The ozone is also distributed through vertical transport in tropical convection followed by lateral transport on isentropic surfaces.

Since the TAR, developing countries have begun reducing emissions from mobile sources though stricter standards. As a result, the A2 scenario should be considered a pessimistic scenario for emissions of ozone precursors. Alternate projections of the evolution of these precursors have been developed with the Regional Air Pollution Information and Simulation (RAINS) model (Amann et al., 2004). It is more likely that the emissions will be consistent with source strengths permitted under the Current Legislation (CLE) scenario, but it also possible that emissions may be reduced more aggressively under a Maximum Feasible Reduction (MFR) scenario. The concentrations of ozone and methane have been simulated for the MFR, CLE, and A2 scenarios for the period 2000 through 2030 using an ensemble of twenty-six chemical transport models (Stevenson et al., 2006). The changes in NO_x emissions for these three scenarios are –27%, +12%, and +55% relative to year 2000. The corresponding changes in ensemble-mean burdens in tropospheric O_3 are –5%, +6%, and +18% for the MFR, CLE, and A2 scenarios, respectively. There are substantial inter-model differences of order ±25% in these results. The ozone decreases throughout the troposphere in the MFR scenario, but the zonal annual-mean concentrations increase by up to 6 ppbv for the CLE scenario and by typically 6 to 10 ppbv in the A2 scenario (Figure 10.4.6).

[INSERT FIGURE 10.4.6 HERE].

The radiative forcing by the combination of ozone and methane changes by –0.05, 0.18, and 0.30 W m^{-2} for these three cases. These results indicate that the growth in tropospheric ozone between 2000 and 2030 could be reduced or reversed depending on emissions controls.

The major issues in the fidelity of these simulations for future tropospheric ozone are the sensitivities to the representation of the stratospheric production, destruction, and transport of O_3 and the exchange of species between the stratosphere and troposphere. Few of the models include the effects of non-methane hydrocarbons (NMHCs), and the sign of the effects of NMHCs on O_3 are not consistent among the models that do (Hauglustaine and Brasseur, 2001; Grenfell et al., 2003).

The effect of more stratosphere-troposphere exchange (STE) in response to climate change increases the concentrations of O_3 in the upper troposphere due to the much greater concentrations of O_3 in the lower stratosphere than the upper troposphere. While the sign of the effect is consistent in recent simulations, the magnitude of the change in STE and its effects on O_3 are very model dependent. In a simulation forced by the SRES A1FI scenario, Collins et al. (2003) find that the downward flux of O_3 increases by 37% from the 1990s to the 2090s. As a result, the concentration of O_3 in the upper troposphere at mid-latitudes increases by 5 to 15%. However, Sudo et al. (2003) and Zeng et al. (2003) predict that STE increases by 80% by 2100. The increase in STE is driven by increases in the descending branches of the Brewer-Dobson circulation at mid-latitudes. The effects of the enhanced STE are sensitive to the simulation of processes in the stratosphere, including the effects of lower temperatures and the evolution of chlorine, bromine, and NO_x concentrations. Since the greenhouse effect (GHE) of O_3 is largest in the upper troposphere, the treatment of STE remains a significant source of uncertainty in the calculation of the total GHE of tropospheric O_3.

The effects of climate change, in particular increased tropospheric temperatures and water vapour, tend to offset some of the increase in O_3 driven by emissions. For example, Stevenson et al. (2000) find that the higher water vapour offsets the increase in O_3 by 17%. The water vapour both decelerates the chemical production and accelerates the chemical destruction of O_3. The photochemical production depends on the concentrations of NO_y, and the additional water vapour causes a larger fraction of NO_y to be converted to HNO_3, which can be efficiently removed from the atmosphere in precipitation (Grewe et al., 2001). The vapour also increases the concentrations of OH through reaction with $O(^1D)$, and the removal of $O(^1D)$ from the atmosphere slows the formation of O_3. The increased concentrations of OH and the increased rates of CH_4 oxidation with higher temperature further reduce the lifetime of tropospheric CH_4 by 12%.

Recent measurements indicate that methane growth rates have declined (see Chapter 2, Executive Summary). The observed rate of increase of 0.8 ppb yr^{-2} for the period 1999 to 2004 is considerably less than the rate of 6 ppb yr^{-2} assumed in all the SRES scenarios for the period 1990 to 2000 (Nakicenovic and Swart, 2000, or Appendix II of the TAR WG1). Recent studies (Dentener et al., 2005) have considered lower emission scenarios (see above) that take account of new pollution-control techniques adopted in major developing countries. In the "Current Legislation" scenario, emissions of CH_4 are comparable to the B2 scenario and increase from 340 Tg yr^{-1} in 2000 to 450 Tg yr^{-1} in 2030. The CH_4 concentrations increase from 1750 ppbv in 2000 to between 2090 and 2200 ppbv in 2030 under this scenario. In the "Maximum Feasible Reduction" scenario, the emissions are sufficiently low that the concentrations in 2030 are unchanged at 1750 ppbv. Under these conditions, the changes in radiative forcing by methane between the 1990s and 2020s are less than 0.01 W m^{-2}.

Current understanding of the magnitude and variation of methane sources and sinks is covered in Chapter 7 Section 7.4, where it is noted that there are substantial uncertainties though the modelling has progressed. There is some evidence for a coupling between climate and wetland emissions. For example, calculations using atmospheric concentrations and small-scale emission measurements as input differ by 60% (Shindell and Schmidt, 2004). The original SRES scenarios for gases other than CO_2 only considered changes in direct anthropogenic emissions, However, the concurrent changes in natural sources of methane are now being estimated to first order using simple models of the biosphere coupled to AOGCMs. Simulations of the response of wetlands to climate change from doubling CO_2 show that wetland emissions increase by 78%. (Shindell and Schmidt, 2004). Most of this effect is caused by growth in the flux of methane from existing tropical wetlands. The increase is equivalent to approximately 20% of current inventories and would contribute an additional 430 ppbv to atmospheric concentrations.

10.4.4 Simulations of Future Evolution of Major Aerosol Species

The time-dependent evolution of major aerosol species and the interaction of these species with climate represent some of the major sources of uncertainty in projections of climate change. An increasing number of AOGCMs have included multiple types of tropospheric aerosols including sulphates, nitrates, black and organic carbon, sea salt, and soil dust. Of the twenty-three models represented in the multi-model ensemble of climate-change simulations for IPCC AR4, ten include other tropospheric species besides sulphates. Of these, seven have the non-sulphate species represented with parameterizations that interact with the remainder of the model physics. Nitrates are treated in just two of the models in the ensemble. Recent projections of nitrate and sulphate loading under the SRES A2 scenario suggest that forcing by nitrates may exceed forcing by sulphates by the end of the 21st century (Adams et al., 2001). This result is of course strongly dependent upon the evolution of precursor emissions for these aerosol species.

The black and organic carbon aerosols in the atmosphere include a very complex system of primary organic aerosols (POA) and secondary organic aerosols (SOA), which are formed by oxidation of biogenic volatile organic compounds. The models used for climate projections typically use highly simplified bulk parameterizations for POA and SOA. More detailed parameterizations for the formation of SOA that trace oxidation pathways have only recently been developed and used to estimate the direct radiative forcing by SOA for present-day conditions (Chung and Seinfeld, 2002). The forcing by SOA is an emerging issue for simulations of present-day and future climate since the rate of chemical formation of SOA may be 60% or more of the emissions rate for primary carbonaceous aerosols (Kanakidou et al., 2005). In addition, two-way coupling between reactive chemistry and tropospheric aerosols has not been explored comprehensively in climate-change simulations. Unified models that treat tropospheric ozone-NO_x-hydrocarbon chemistry, aerosol formation, heterogeneous processes in clouds and on aerosols, and gas-phase photolysis have been developed and applied to the current climate (Liao et al., 2003). However, to date these unified models have not yet been used extensively to study the evolution of the chemical state of the atmosphere under future scenarios.

The interaction of soil dust with climate is under active investigation. Whether emissions of soil dust aerosols increase or decrease in response to changes in atmospheric state and circulation is still unresolved (Tegen et al., 2004a). Several recent studies have suggested that the total surface area where dust can be mobilized will decrease in a warmer climate with higher concentrations of carbon dioxide (e.g., Harrison et al., 2001). The interaction of and net effects of reductions in dust emissions from natural sources combined with land-use change could potentially be significant but have not been systematically modelled as part of climate-change assessment.

Uncertainty regarding the scenario simulations is compounded by inherently unpredictable forcings from future volcanic eruptions and solar variability. The eruptions that produce climatologically significant forcing represent just the extremes of global volcanic activity (Naveau and Ammann, 2005). Global simulations are not yet available for the effects of future volcanic eruptions modelled stochastically using the statistical properties of prior eruptions.

10.5 Quantifying the Range of Climate Change Projections

10.5.1 Sources of Uncertainty and Hierarchy of Models

Uncertainty in predictions of anthropogenic climate change is injected at all stages of the modelling process described in Section 10.1. The specification of future emissions of greenhouse gases, aerosols and their precursors is uncertain (e.g., Nakicenovic and Swart, 2000). Following Figure 10.1.1 it is then necessary to convert these emissions into concentrations of radiatively active species, calculate the associated forcing and predict the response of climate system variables such as surface temperature and precipitation, by simulating the relevant physical and biogeochemical processes. At each step uncertainty in the true signal of climate change is introduced both by errors in the representation of processes in models (e.g., Palmer et al., 2005) or by noise associated with unforced internal climate variability (e.g., Selten et al., 2004). The effects of internal variability can be quantified by running models many times from different initial conditions, assuming that simulated internal variability is consistent with that observed in the real world. The effects of uncertainty in our knowledge of Earth system processes can be partially quantified by constructing ensembles of models which sample different parameterisations of these processes. However, some processes may be missing from the set of available models, and the range of options for the parameterisation of other processes may share common systematic biases. Such structural inadequacies imply that distributions of changes obtained from ensembles of models are themselves subject to uncertainty (Smith, 2002). Therefore, distributions of future climate responses from ensemble simulations are dependent upon the quality of the available models and the range of process uncertainties which they explore. These distributions may be modified to reflect observational constraints expressed through metrics of the agreement between the observed historical climate and the simulations of individual ensemble members, for example through Bayesian methods (see Chapter 9, Appendix 9.B). In this case, the choice of observations and their associated errors introduces further sources of uncertainty, and the distributions would be wider were uncertainty due to structural model errors to be accounted for.

A spectrum or hierarchy of models of varying complexity has been developed (Claussen et al., 2002; Stocker and Knutti, 2003) to assess the range of future changes consistent with our understanding of known uncertainties. Simple climate models (SCMs) typically represent the ocean-atmosphere system as a set of global or hemispheric boxes, predicting global surface temperature using an energy balance equation, a prescribed value of climate sensitivity and a basic representation of ocean heat uptake (see Chapter 8, Section 8.8.2). Their role is to perform comprehensive analyses of the interactions between global variables, based on prior estimates of uncertainty in their controlling parameters obtained from observations, expert judgement and from tuning to complex models. By coupling SCMs to simple models of biogeochemical cycles they can be used to extrapolate the results of AOGCM simulations to a wide range of alternative forcing scenarios (e.g., Wigley and Raper, 2001, see Chapter 10, Section 10.5.3).

Compared to SCMs, Earth system models of intermediate complexity (EMICs) include more of the processes simulated in AOGCMs, but in a less detailed, more highly parameterised form (see Chapter 8, Section 8.8.3), and at coarser resolution. Consequently, EMICS are not suitable for quantifying uncertainties in regional climate change or extreme events, however they can be used to investigate the large scale effects of coupling between multiple Earth system components in large ensembles or long simulations (e.g., Forest et al, 2002), which is not yet possible with AOGCMs. To this end, a number of EMICs include biological and geochemical modules such as vegetation dynamics, the terrestrial and ocean carbon cycles and atmospheric chemistry (Claussen et al., 2002), filling a gap in the spectrum of models between AOGCMs and SCMs. Thorough sampling of parameter space is computationally feasible for some EMICs (e.g., Stocker and Schmittner, 1997; Forest et al., 2002; Knutti et al., 2002), as for SCMs (Wigley and Raper, 2001), and is used to obtain probabilistic projections (see Section 10.5.2). In some EMICs climate sensitivity is an adjustable parameter, as in SCMs. In other EMICs climate sensitivity is dependent on multiple model parameters, as in AOGCMs. Probabilistic estimates of climate sensitivity and transient climate response from SCMs and EMICs are assessed in Chapter 9, Section 9.6 and compared with estimates from AOGCMs in Box 10.2.

The high resolution and detailed parameterisations in AOGCMs make them the only modelling tools capable of realistic simulation of internal variability (see Chapter 8, Section 8.4), extreme events (see Chapter 8, Section 8.5), and the complex interactions which drive global and regional climate change feedbacks (Boer and Yu, 2003a; Bony and Dufresne, 2005; Bony et al., 2006; Soden and Held, 2006).

Given that ocean dynamics plays a significant role in determining regional feedbacks (Boer and Yu, 2003b), it is clear that quantification of regional uncertainties in time-dependent climate change requires multi-model ensemble simulations with AOGCMs containing a full, three-dimensional dynamic ocean component. However, downscaling methods (see Chapter 11) are required to obtain credible information at spatial scales near or below the AOGCM grid scale (125–400 km in the AR4 AOGCMs, see Chapter 8, Table 8.2.1).

Uncertainty ranges for 21st century climate change obtained from the AR4 multi-model ensemble of AOGCM simulations and from EMIC simulations are discussed in Section 10.5.2, considering selected forcing scenarios. Section 10.5.3 presents the role of uncertainties in emissions, using SCMs and EMICs calibrated to the multi-model ensemble results. Developments since the TAR in the use of AOGCM ensembles to sample uncertainties and estimate probabilities for climate change are summarised in Section 10.5.4.

10.5.2 Range of Responses from Different Models

10.5.2.1 Comprehensive AOGCMs

The way a climate model responds to changes in external forcing, such as an increase in anthropogenic GHGs, is characterized by two standard measures: (1) *equilibrium climate sensitivity* (the equilibrium change in global surface temperature following a doubling of the atmospheric equivalent CO_2 concentration, see glossary), and (2) *transient climate response* (TCR, the change in global surface temperature in a global coupled climate model in a 1% per year CO_2 increase experiment at the time of CO_2 doubling, see glossary). The first measure provides an indication of feedbacks mainly residing in the atmospheric model but also in the land surface and sea ice components, and the latter quantifies the response of the fully coupled climate system including aspects of transient ocean heat uptake (e.g., Sokolov et al., 2003). These two measures have become standard metrics of climate model response to understand how an AOGCM will react to more complicated forcings in scenario simulations.

Historically, the equilibrium climate sensitivity has been given being likely in the range from 1.5°C to 4.5°C. This range has also been reported in the TAR with no indication of a probability distribution within this range. However, considerable recent work has addressed the range of equilibrium climate sensitivity, and attempted to assign probabilities of climate sensitivity.

Equilibrium climate sensitivity and TCR are not independent (Figure 10.5.1a). A large ensemble of the Bern2.5D EMIC has been used to explore the relationship of TCR and equilibrium sensitivity over a wide range of ocean heat uptake parameterizations (Knutti et al., 2005). Good agreement with the available results from AOGCMs is found, and the EMIC covers almost the entire range of structurally different models. Similarly, the percent change in precipitation is closely related to the equilibrium climate sensitivity for the current generation of AOGCMs (Figure 10.5.1b), with values from the current models falling within the range of the models from the TAR . Figure 10.5.1c shows the percent change of globally averaged precipitation at time of CO_2 doubling from 1% per year transient CO_2 increase experiments with AOGCMs as a function of TCR suggesting a broadly positive correlation between these two quantities similar to that for equilibrium climate sensitivity, with these values from the new models also falling within the range of the previous generation of AOGCMs assessed in the TAR.

[INSERT FIGURE 10.5.1 HERE]

Fitting normal distributions, the resulting 5–95% uncertainty range for equilibrium climate sensitivity from the AOGCMs is approximately 2.0°C–4.4°C and that for TCR 1.2°C–2.4°C (using the method of Räisänen, 2005b). The median for climate sensitivity is 3.2°C and that for TCR 1.8°C. These numbers are practically the same for both the normal and the log-normal distribution (see Box 10.2). Therefore, assuming that current global coupled climate models cover the full range of uncertainty and assuming a certain shape of distribution, the "range" of equilibrium climate sensitivity can be modified from an equal probability of any value from 1.5°C to 4.5°C, to a 90% confidence interval of 2.0°C to 4.4°C, with a median of 3.2°C. The assumption of a (log)-normal fit is not well supported from the limited sample of AOGCM data. Also, the AOGCMs represent an 'ensemble of opportunity' and are by design not sampled in a random way. However, most studies aiming to constrain climate sensitivity by observations do indeed indicate a log-normal distribution of climate sensitivity and an approximately normal distribution of the uncertainty in future warming and thus TCR (see Box 10.2). On the other hand, there is a consensus from most studies cited above using observational constraints (see Section 10.5.4.) that the upper limit of climate sensitivity is uncertain, with a substantial probability for sensitivity above 4.5°C, and that the current AOGCMs therefore do not cover the full possible range of sensitivities.

The nonlinear relationship between TCR and equilibrium climate sensitivity shown in Figure 10.5.1a also indicates that on time scales well short of equilibrium, the model's transient climate response is not particularly sensitive to model's climate sensitivity. The implication is that transient climate change is better constrained than the equilibrium climate sensitivity, i.e., models with different sensitivity might still show good agreement for projections on decadal timescales. Therefore, in the absence of unusual solar or volcanic activity, climate change is well constrained for the coming few decades. The reasons for that are that differences in some feedbacks will only become important on long timescales (see also Section 10.5.4.5, Figure 10.5.6) and that over the next few decades, about half of the projected warming is the commitment warming already caused by well known changes in radiative forcing in the past (see Section 10.7).

These types of metrics provide information on the possible range and maximum likelihood, and can be related to PDFs from perturbed physics ensembles discussed in Section 10.5.4.

Different treatments of boundary layer cloud processes appear to be important for climate sensitivity (see Chapter 8, Section 8.6). In models that have been contributed to CFMIP (AGCMs run to equilibrium with a slab ocean) it was found that the differences in global equilibrium climate sensitivity are most influenced by two differing geographic areas, the sea ice region (where albedo feedback is dominant) and the actual area covered by model simulated low stratus clouds (see Chapter 8, Bony and Dufresne, 2005). Improvements in simulation of low stratus for the current climate in some models have dramatically altered the contribution of changes in low cloud amounts to the climate sensitivity (Webb et al., 2006).

Further indications of equilibrium climate sensitivity can be obtained from other forcing simulations, such as LGM, Pinatubo, or the 20th century, to better constrain model sensitivity (see Chapter 9). A summary of all of the estimates of equilibrium climate sensitivity is given in Box 10.2.

10.5.2.2 Earth system models of intermediate complexity

Over the last few years a range of climate models has been developed that are dynamically simpler and of lower resolution than comprehensive AOGCMs, although they might well be more "complete" in terms of climate system components that are included. The class of such models, usually referred to as Earth System Models of Intermediate Complexity (EMICs, Claussen et al., 2002), is very heterogeneous ranging from zonally averaged ocean models coupled to energy balance models (Stocker et al., 1992a), or coupled to statistical-dynamical models of the atmosphere (Petoukhov et al., 2000), to low resolution 3-dimensional ocean models, coupled to energy balance or simple dynamical models of the atmosphere (Opsteegh et al., 1998; Edwards and Marsh, 2005). Some EMICs have a radiation code and prescribe greenhouse gases, while others use simplified equations to project radiative forcing from projected concentrations and abundances (Joos et al., 2001, see Chapter 2 and IPCC TAR, 2001, Appendix II, Table II.3.11). Compared to comprehensive models, EMICs place hardly any computational constraint, and therefore many simulations can be performed. This allows for the creation of large ensembles, or the systematic exploration of long-term changes many centuries hence. However, because of the reduced complexity, only results on the largest scales, continental to global, are to be interpreted (Stocker and Knutti, 2003). Chapter 8, Table 8.8.1 lists all EMICs used in this section, including their components and resolution.

A set of simulations is used to compare EMICs with AOGCMs for the SRES A1B with stable atmospheric concentrations after year 2100 (see Section 10.7.2). For global mean temperature and sea level, the EMICs generally reproduce the AOGCM behaviour quite well. Two of the EMICs have values for climate sensitivity and transient response below the AOGCM range. However, climate sensitivity is a tuneable parameter in some EMICs, and no attempt was made here to match the range of response of the AOGCMs. The transient reduction of the MOC in most EMICs is also similar to the AOGCMs (see also Sections 10.3.4 and 10.7.2, Figure 10.7.3), providing support that this class of models can be used for both long-term commitment projections (see Section 10.7) and probabilistic projections involving hundreds to thousands of simulations (see Section 10.5.4.5). If the forcing is strong enough, and lasts long enough (e.g., 4 × CO_2, not shown), a complete and irreversible collapse of the MOC can be induced in a few models. This is in line with earlier results using EMICs (Stocker and Schmittner, 1997; Rahmstorf and Ganopolski, 1999), or a coupled model (Stouffer and Manabe, 1999).

10.5.3 Global Mean Responses from Different Scenarios

10.5.3.1 SRES scenarios from simple models and sensitivity to uncertainties

The TAR projections with a simple climate model presented a range of warming over the 21st century for 35 SRES scenarios. SRES emission scenarios assume that no climate policies are implemented (Nakicenovic and Swart, 2000). The construction of the TAR Chapter 9, Figure 9.14 was pragmatic. It used a simple model tuned to AOGCMs that had a climate sensitivity within the long-standing range of 1.5–4.5°C (e.g., Charney, 1979, and stated in earlier IPCC Assessment Reports). Models with climate sensitivity outside that range were discussed in the text and allowed the statement that the presented range was not the extreme range indicated by AOGCMs. The figure was based on a single anthropogenic-forcing estimate for 1750 to 2000, which is well within the range of values recommended by TAR Chapter 6, and is also consistent with that deduced from model simulations and the observed temperature record (TAR Chapter 12.). To be consistent with TAR Chapter 3, climate feedbacks on the carbon cycle were included. The resulting range of global mean temperature change from 1990 to 2100 given by the full set of SRES scenarios was 1.4 to 5.8°C.

Since the TAR several studies have examined the TAR projections and attempted probabilistic assessments. Allen et al. (2000) show that the forcing and simple climate model tunings used in the TAR give projections that are in agreement with the observationally constrained probabilistic forecast, reported in TAR Chapter 12, stating that under the IS92a scenario anthropogenic warming is likely to lie in the range 0.1° to 0.2°C per decade over the next few decades.

As noted by Moss and Schneider (2000), giving only a range of warming results is potentially misleading unless some guidance is given as to what the range means in probabilistic terms. Wigley and Raper (2001) interpret the warming range in probabilistic terms, accounting for uncertainties in emissions, the climate sensitivity, the carbon cycle, ocean mixing, and aerosol forcing. They give a 90% probability interval for 1990 to 2100 warming of 1.7° to 4°C. As pointed out by Wigley and Raper (2001), such results are only as realistic as the assumptions upon which they are based. Key assumptions in this study were: that each SRES scenario was equally likely, that 1.5° to 4.5° corresponds to the 90% confidence interval for the climate sensitivity, and that carbon cycle feedback uncertainties can be characterised by the full uncertainty range of abundance in 2100 of 490 to 1260 ppm given in the TAR. The aerosol probability density function (PDF) was based on the uncertainty estimates given in the TAR together with constraints based on fitting the simple climate model to observed global- and hemispheric-mean temperatures.

The most controversial assumption in the Wigley and Raper (2001) probabilistic assessment was the assumption that each SRES scenario was equally likely (see AR4 WGII Chapter 2, Section 2.2.3.3). The Special Report on Emissions Scenarios (Nakicenovic and Swart, 2000) states that *'No judgment is offered in this report as to the preference for any of the scenarios and they are not assigned probabilities of occurrence, neither must they be interpreted as policy recommendations'*. Problems in trying to nevertheless assign probabilities to the scenarios include, for example, the fact that the individual scenarios are clearly not independent. Furthermore, the source of an overriding objection to assigning probabilities maybe found in the following quote from the SRES report: '*As required by the Terms of Reference, the scenarios in this report do not include additional climate initiatives, which means that no scenarios are included that explicitly assume implementation of the United Nations Framework Convention for Climate Change (UNFCCC) or the emissions targets of the Kyoto Protocol.*' It clearly follows that the SRES scenarios cannot be regarded as capturing an agreed sense of the range of future options.

Webster et al. (2003) use the probabilistic emissions projections of Webster et al. (2002), which consider present uncertainty in SO_2 emissions, and allow the possibility of continuing increases in SO_2 emissions over the 21st century, as well as the declining emissions consistent with SRES. Since their climate model parameter PDFs were constrained by observations and are mutually dependent the effect of the lower present day aerosol forcing on the projections is not easy to separate, but there is no doubt that their projections tend to be lower where they admit higher and increasing SO_2 emissions.

Irrespective of the question, whether it is possible to assign probabilities to specific emissions scenarios, it is important to distinguish different sources of uncertainties for temperature projections until 2100. Clearly, one major uncertainty is the 'emission uncertainty' which arises from the fact that future greenhouse gas emissions are largely dependent on key socio-economic drivers, technological development and political decisions, which are hardly predictable. On the other hand, the 'response uncertainty' is defined as the range in projections for a particular emission scenario and arises merely from our limited knowledge of how the climate system will react to the anthropogenic perturbations. In the following, all given uncertainty ranges therefore reflect merely the response uncertainty of the climate system and should therefore be seen as conditional on a specific emission scenario.

The following paragraphs describe the construction of the AR4 temperature projections for the 6 illustrative SRES marker scenarios, using the simple climate model tuned to 19 models from the IPCC AR4 data set (see Chapter 8, Section 8.8). These 19 tuned simple model versions have climate sensitivities in the range 1.90°C to 4.96°C. The simple model sensitivities are derived from the full coupled 2x and 4xCO2 1% CO_2 increase per year AOGCM simulations and in some cases differ from the equilibrium slab ocean model sensitivities given in Chapter 8, Table 8.8.1.

The SRES emission scenarios used here, were designed to represent plausible futures assuming that no climate policies will be implemented. This chapter does not analyse any scenarios with explicit climate change mitigation policies. Still, there is a wide variation across these SRES scenarios in terms of anthropogenic emissions, such as those of fossil CO_2, CH_4, and SO_2 (Nakicenovic and Swart, 2000) as shown in the top three panels of Figure 10.5.2.

[INSERT FIGURE 10.5.2 HERE]

As a direct consequence of the different emissions, the projected concentrations vary widely for the 6 illustrative SRES scenarios – see panel rows 4 to 6 in Figure 10.5.2 for the concentrations of the main greenhouse gases, CO_2, CH_4, and N_2O. These results incorporate the effect of carbon cycle uncertainties (see Section 10.4.1), which were not explored in the TAR. Projected methane concentrations are influenced by the temperature-dependent water vapour feedback on the lifetime of methane. In Figure 10.5.2, the plumes of CO_2 concentration reflect high and low carbon cycle feedback settings of the applied simple climate model. Their derivation is described as follows. The carbon cycle model in the SCM used here (MAGICC) includes a number of climate-related carbon cycle feedbacks driven by global-mean temperature. The parameterization of the overall effect of carbon cycle feedbacks is tuned to the more complex and physically realistic carbon cycle models of the C4MIP intercomparison (Friedlingstein and Solomon, 2005, also see Section 10.4). This allows the SCM to produce probabilistic projections of future CO_2 concentration change that are consistent with state-of-the-art carbon cycle model results. Specifically, the C4MIP range of 2100 CO_2 concentrations for the A2 emission scenario is 730 to 1020 ppm, while the simple model results presented here show an uncertainty range from 805 ppm to 1000 ppm. The lower bound of this simple model uncertainty range is the mean minus 1 standard deviation (std) for low carbon cycle feedback settings and the 19 AOGCM tunings, while the upper bound represents the mean plus 1 std for high carbon cycle settings. For comparison, the 90% confidence interval from Wigley and Raper (2001) is 770 to 1090 ppm. The simple model CO_2 concentration projections can be slightly higher than under the C4MIP inter-comparison because the simple model's carbon cycle is driven by the full temperature changes in A2, while the C4MIP values are driven by the component of A2 climate change due to CO_2 alone.

The radiative forcing projections combine anthropogenic and natural, solar and volcanic, forcing as shown in Figure 10.5.2. The forcing plumes reflect primarily the sensitivity of the forcing to carbon cycle uncertainties. Results are based on a forcing of 3.71 W m^{-2} for a doubling of the carbon dioxide concentration. The anthropogenic forcing is based on Chapter 2, Table 2.9.1 but uses a value of –0.8 W m^{-2} for the present day indirect aerosol forcing. Solar forcing for the historical period is prescribed according to Lean et al. (1995) and volcanic forcing according to Ammann et al. (2003). The historic solar forcing series is extended into the future by its average over the most recent 22 years. The volcanic forcing is adjusted to have a zero mean over the past 100 years and is assumed to be zero for the future. In the TAR the anthropogenic forcing was used alone even though the projections started in 1765. There are several advantages of using both natural and anthropogenic forcing for the past. First, this is that this it what was done by most AOGCMs the simple models are emulating. Second, it allows the simulations to be compared with observations and third, the warming commitments accrued over the instrumental period are reflected in the projections. The disadvantage of including natural forcing is that the warming projections in 2100 are dependent to a few tenths of a degree on the necessary assumptions made about the natural forcing. These assumptions include how the natural forcing is projected into the future and whether to reference the volcanic forcing to a past reference period mean value. Also the choice of data set for both solar and volcanic forcing affects the results.

The temperature projections for the six marker scenarios are shown in the bottom panel of Figure 10.5.2. Model results are referenced to the mean of the historical observations (Folland et al., 2001; Jones et al., 2001; Jones and Moberg, 2003) over the 1980 to 2000 period and the corresponding observed temperature anomalies are shown for comparison. The inner (darker) plumes show the ±1 standard deviation uncertainty due to the 19 model tunings and the outer (lighter) plumes show results for the corresponding high and low carbon cycle settings. Note the asymmetry in the carbon cycle uncertainty causes global mean temperature projections to be skewed towards higher warming.

The results from this section for warming at the end of the 21st century are summarized in Figure 10.5.3. Reflecting the importance of separating 'emission' and 'response' uncertainty, this figure shows global mean surface temperature projections in 2100 ('response') plotted against emissions, quantified by the cumulative GHG emissions over the period 1990 to 2100. The 100-year Global Warming Potentials as provided by the IPCC in its Second Assessment Report are used for the aggregation of CO_2, CH_4, N_2O, HFCs, PFCs and SF_6 emissions, consistent with the guidelines for national communications under the United Nations Framework Convention on Climate Change. For the 6 illustrative SRES non-mitigation marker scenarios, the cumulative global GHG emissions from 1990 to 2100 vary between 4.9 (B1) and 10.2 (A1FI) Teratons of CO_2 equivalent (1 Tera ton = 10^{18} grams). The relationship between cumulative GHG emissions and 2100 temperatures is influenced by differences in gas-by-gas emission ratios, SO_2 emissions and different timing of emissions across the different SRES scenarios (see Figure 10.5.3).

[INSERT FIGURE 10.5.3 HERE]

The comparison of the simple model results with the individual AOGCM results (black dots on Figure 10.5.3) is not straightforward since the number of AOGCM simulations differs for each scenario (AOGCMs were run for B1, A1B and A2 scenarios). However, analysis of the multi-model AOGCM ensemble shows that for a given subset of models, the fractional uncertainty (the standard deviation of global temperature increase across models divided by the mean temperature increase) is roughly independent of the scenario and time after the year 2050, and is estimated to be approximately 20% for A1B where the largest number of models is available. These 'mean ±20%' uncertainty ranges for the AOGCM results span from 1.5°C to 2.2°C (B1), 2.2°C to 3.3°C (A1B) and 2.7°C to 4.0°C (A2) – see vertical black lines in Figure 10.5.3.

The largest difference between the AOGCM range and the simple model range is for scenario A2 and is because the high sensitivity AOGCM, MIROC 3.2 (hires), did not run scenario A2. This leads to a slightly smaller mean and smaller standard deviation for the original AOGCM data compared to the simple climate model results, which consistently used the 19 AOGCM tunings for all emission scenarios. There is also the minor effect that the AOGCM results are averages over the 2090–2100 period, not for the year 2100. Apart from these disparities and assuming that the simple climate model successfully emulates the AOGCMs, the differences between the AOGCM and the simple model results should reflect differences in the forcing. The simple climate model used harmonized forcing whereas this is not the case with the AOGCMs (see Section 10.2). Another possible source of difference is that more than half of the analysed AOGCMs did not include indirect aerosol cooling effects. Since SO_2 emissions in the SRES scenarios with AOGCM results are lower than present day in 2100, the simple climate model suggests a more pronounced warming between present times (characterized by relatively high aerosol cooling) and the end of the century (with relatively low aerosol cooling).

Considering only the mean of the simple climate model results with medium carbon cycle settings, the global mean temperature change between 1990 and 2100 for the lower SRES emission scenario B1 is 2.0°C. For a higher emission scenario, for example SRES A2 scenario, the global mean temperature is projected to rise by 3.8°C between 1990 and 2100 (see 'SCM mean' uncertainty bars in Figure 10.5.3). This clear difference in projected mean warming highlights the importance of assessing different emission scenarios separately. As mentioned above, the 'response uncertainty' is defined as the range in projections for a particular emission scenario. For the A2 emission scenario, the temperature change projections with the simple climate model span a ±1 standard deviation range of about 1.5°C, from 3.1 to 4.6°C in 2100 above 1980–2000 levels. If carbon cycle feedbacks are considered to be low, the lower end of this range decreases only slightly to 3.0°C. For the higher carbon cycle feedback settings, the upper bound of the ±1 standard deviation range increases to 5.0°C. For lower emission scenarios this uncertainty range is smaller. For example, the B1 scenario projections span a range of 1.3°C, from 1.5°C to 2.8°C, including carbon cycle uncertainties (see light shaded bars in Figure 10.5.3). The corresponding results for the medium emission scenario A1B are 2.3°C to 4.1°C, and for the higher emission scenario A1FI, they are 3.5°C to 5.8°C. Note that these uncertainty ranges are not the minimum to maximum bounds of the projected warming across all simple climate model runs, which are higher, namely 2.6°C to 6.3°C for the A2 scenario and 1.3°C to 3.7°C for the B1 scenario (not shown).

The simple climate model results presented here are a sensitivity study with different model tunings and carbon cycle feedback parameters. Note that forcing uncertainties have not been assessed. Also note that the AOGCM model results available for simple climate model tuning may not span the full range of possible climate response. For example, studies that constrain forecasts based on model fits to historic or present day observations generally allow for a somewhat wider 'response uncertainty' (see Section 10.5.4). The concatenation of all such uncertainties would require a probabilistic approach because the extreme ranges have low probability.

10.5.4 Sampling Uncertainty and Estimating Probabilities

Uncertainty in the response of an AOGCM arises from the effects of internal variability, which can be sampled in isolation by creating ensembles of simulations of a single model using alternative initial conditions, and from modelling uncertainties, which arise from errors introduced by the discretisation of the equations of motion on a finite resolution grid, and errors in the parameterisation of sub-grid scale processes (radiative transfer, cloud formation, convection etc). Modelling uncertainties are manifested in alternative structural choices (for example, choices of resolution and the basic physical assumptions on which parameterisations are based), and in the values of poorly-constrained parameters within parameterisation schemes. Ensemble approaches are used to quantify the effects of uncertainties arising from variations in model structure and parameter settings. These are assessed in Sections 10.5.4.1–10.5.4.3, followed by a discussion of observational constraints in Section 10.5.4.4 and methods used to obtain probabilistic predictions in Sections 10.5.4.5 and 10.5.4.6.

While ensemble predictions carried out to date give a wide range of responses, they do not sample all possible sources of modelling uncertainty. For example, the AR4 multi-model ensemble relies on specified concentrations of CO_2, thus neglecting uncertainties in carbon cycle feedbacks (see Section 10.4.1), though this can be partially addressed by using less detailed models to extrapolate the AOGCM results (see Section 10.5.3). More generally, the set of available models may share fundamental inadequacies, the effects of which cannot be quantified (Kennedy and O'Hagan, 2001). For example, climate models currently implement a restricted approach to the parameterisation of sub-grid scale processes, using deterministic bulk formulae coupled to the resolved flow exclusively at the grid scale. Palmer et al. (2005) argue that the outputs of parameterisation schemes should be sampled from statistical distributions consistent with a range of possible sub-grid scale states, following a stochastic approach which has been tried in numerical weather forecasting (e.g., Buizza et al., 1999; Palmer, 2001). The potential for missing or inadequately parameterised processes to broaden the simulated range of future changes is not clear, however, this is an important caveat on the results discussed below.

10.5.4.1 The multi-model ensemble approach

The use of ensembles of AOGCMs developed at different modelling centres has become established in climate prediction/projection on both seasonal to interannual and centennial time scales. To the extent that simulation errors in different AOGCMs are independent, the mean of the ensemble can be expected to outperform individual ensemble members, thus providing an improved "best estimate" forecast. Results show this to be the case, both in verification of seasonal forecasts (Palmer et al., 2004; Hagedorn et al., 2005) and of the present day climate from long term simulations, both in the AR4 multi-model ensemble (see Chapter 8, Section 8.3) and earlier ensembles (Lambert and Boer, 2001). By sampling modelling uncertainties, ensembles of AOGCMs should provide an improved basis for probabilistic projections compared with ensembles of a single model sampling only uncertainty in the initial state (Palmer et al., 2005). Probabilistic verification of future climate change projections is not possible, however, Räisänen and Palmer (2001) used a "perfect model approach" (treating one member of an ensemble as truth and predicting its response using the other members) to show that the hypothetical economic costs associated with climate events can be reduced by calculating the probability of the event across the ensemble, rather than using a deterministic prediction from an individual ensemble member.

A strength of multi-model ensembles is that each member is subjected to careful testing in order, for example, to achieve a plausible and stable control simulation. However, members of a multi-model ensemble still share common systematic errors (Lambert and Boer, 2001), and cannot span the full range of possible model configurations due to resource constraints.

10.5.4.2 Perturbed physics ensembles

The AOGCMs featured in Section 10.5.2 are built by selecting components from a pool of alternative parameterisations, each based on a given set of physical assumptions and including a number of uncertain parameters. A comprehensive approach to quantifying the range of predictions consistent with these components would be to construct very large ensembles with systematic sampling of multiple combinations of options for parameterisation schemes and parameter values. SCMs and EMICs have recently adopted such an approach (Wigley and Raper, 2001; Knutti et al., 2002) and Murphy et al. (2004) and Stainforth et al. (2005) describe the first steps in this direction using AOGCMs, constructing large ensembles by perturbing poorly constrained surface and atmospheric parameters in the atmospheric component of HadCM3 coupled to a mixed layer ocean. These experiments quantify the range of equilibrium responses to doubled CO_2 consistent with uncertain parameters in a single GCM, however, they do not sample uncertainties associated with changes in ocean circulation, or "structural" model perturbations (see above). Murphy et al. (2004) perturbed 29 parameters one at a time, assuming that effects of individual parameters were additive but making a simple allowance for additional uncertainty introduced by non-linear interactions. They found a probability distribution for climate sensitivity with a 5–95% range of 2.4–5.4°C when weighting the models with a broadly-based metric of the agreement between simulated and observed climatology, cf 1.9–5.3°C when all model versions are assumed equally reliable (Figure 10.5.4).

[INSERT FIGURE 10.5.4 HERE]

Stainforth et al. (2005) deployed a distributed computing approach (Allen, 1999) to run a very large ensemble of 2578 simulations sampling combinations of high, intermediate and low values of six parameters known to affect climate sensitivity. They found climate sensitivities ranging from 2–11°C, with 4.2% of model versions giving values exceeding 8°C, and showed that the high sensitivity models could not be ruled out, based on a comparison with surface annual mean climatology. By utilizing multivariate linear relationships between climate sensitivity and spatial fields of several present day observables the 5–95% range of climate sensitivity is estimated at 2.2–6.8°C from the same dataset (Piani et al., 2005, Figure 10.5.4). Furthermore, in this ensemble, Knutti et al. (2006) find a strong relationship between climate sensitivity and the amplitude of the seasonal cycle in surface temperature in the present day simulations. Most of the simulations with high sensitivities overestimate the observed amplitude. Using this technique, the 5–95% range of climate sensitivity is estimated at 1.5–6.4°C (Figure 10.5.4). The differences between the PDFs in Figure 10.5.4, which are all based on the same climate model, reflect uncertainties in methodology arising from choices of uncertain parameters, their expert-specified prior distributions, and alternative applications of observational constraints. In addition, further work is needed to determine how to account for structural model errors when calculating the relative likelihoods of alternative model versions (Smith, 2002; Goldstein and Rougier, 2005).

Annan et al. (2005a) use an ensemble Kalman Filter technique to obtain uncertainty ranges for model parameters in an EMIC subject to the constraint of minimising simulation errors with respect to a set of climatological observations. Hargreaves and Annan (2006) use this method to explore the range of responses of the Atlantic meridional overturning circulation to increasing CO_2, finding that the risk of a collapse in the circulation depends on the set of observations to which the EMIC parameters are tuned. Chapter 9, Section 9.6.2.3 compares perturbed physics studies of the link between climate sensitivity and cooling at the Last Glacial Maximum based on alternative models and parameter perturbation techniques (Annan et al., 2005b; Schneider von Deimling et al., 2006)

10.5.4.3 Diagnosing drivers of uncertainty from ensemble results

Figure 10.5.5a shows the agreement between annual changes simulated by members of the AR4 multi-model ensemble for 2080–2099 relative to 1980–1999 for the A1B scenario, calculated as in Räisänen (2001). The agreement increases with spatial scale for surface temperature and (especially) precipitation. Differences in model formulation are the dominant contributor to ensemble spread, though the role of internal variability increases at smaller scales (Figure 10.5.5b). These conclusions are consistent with results obtained from the CMIP2 ensemble by Räisänen (2001), and reported in the TAR. However the agreement between AR4 ensemble members is slightly higher compared with the CMIP2 ensemble, and internal variability explains a smaller fraction of the ensemble spread. This is consistent with the larger forcing and responses in the A1B scenario at 2080–2099 compared to the transient response to doubled CO_2 considered by Räisänen (2001), though the use of an updated set of models may also contribute. For seasonal changes, internal variability is found to be comparable with model differences as a source of uncertainty in local precipitation and sea level pressure changes (though not for surface temperature), in both multi-model and perturbed physics ensembles (Räisänen, 2001; Murphy et al., 2004).

Wang and Swail (2006b) examine the relative importance of internal variability, differences in radiative forcing and model differences in explaining the transient response of ocean wave height using three AOGCMs each run for three plausible forcing scenarios, finding model differences to be the largest source of uncertainty in the simulated changes.

[INSERT FIGURE 10.5.5 HERE]

Selten et al. (2004) report a 62 member initial condition ensemble of simulations of 1940–2080 including natural and anthropogenic forcings. They find an individual member which reproduces the observed trend in the NAO over the past few decades, but no trend in the ensemble-mean, and suggest that the observed change can be explained through internal variability associated with a mode driven by increases in precipitation over the tropical Indian Ocean. Terray et al. (2004) find sensitivity to grid resolution in simulated changes in the residence of the positive phase of the NAO. The coupled ocean-atmosphere version of the ARPEGE model shows small increases in response to SRES A2 and B2 forcing, whereas larger increases are found when SST changes prescribed from the coupled version are used to drive a version of the atmosphere model with enhanced resolution over the North Atlantic and Europe (Gibelin and Déqué, 2003).

Collins et al. (2006a) report a perturbed physics ensemble simulation of the transient response to increasing CO_2, created by coupling the ocean component of HadCM3 to 17 versions of the atmospheric component. These are designed to sample a wide range of multiple parameter perturbations and climate sensitivities, subject to the constraint that each version should produce a credible simulation of present day climate. The results (Figure 10.5.6) show a similar range of global mean surface temperature changes to that found in the AR4 multi-model ensemble. These ensembles are complementary in the sense that the AR4 ensemble partially samples ocean physics uncertainties, and the effects of structural variations in atmospheric model components, whereas the perturbed physics ensemble samples parameter uncertainties for a fixed choice of ocean model and fixed structural choices for atmospheric components.

[INSERT FIGURE 10.5.6 HERE]

Soden and Held (2006) find that differences in cloud feedback are the dominant source of uncertainty in the transient response of surface temperature in the AR4 ensemble, as in previous IPCC assessments. Webb et al. (2006) compare equilibrium radiative feedbacks in a 9 member multi-model ensemble against those simulated in a 128 member perturbed physics ensemble with multiple parameter perturbations. They find that the range of climate sensitivities in the multi-model ensemble is explained mainly by uncertainty in the response of shortwave cloud forcing, while the range found in the perturbed physics ensemble is driven more by uncertainties in the response of longwave cloud forcing, with variations in shortwave cloud forcing playing a lesser role. Narrowing uncertainties in cloud feedback is likely to require an improved ability to discriminate between alternative methods of parameterising cloud microphysical properties (Johns et al., 2006; Tsushima et al., 2006). Significant increases in resolution may also be required, in order to resolve more of the atmospheric processes responsible for the observed distribution of cloud and water vapour (Palmer et al., 2005). Given the substantial nature of such challenges, ensemble approaches are likely to remain necessary for the foreseeable future.

10.5.4.4 Observational constraints

A range of observables has been used since the TAR to explore methods for constraining uncertainties in future climate change. These include changes in surface and upper air temperatures and ocean heat content during the past 50–150 years, which have been used in a number of studies with simple climate models, EMICs and AOGCMs to constrain climate sensitivity and transient response in scenarios (see Section 10.5.4.5). Probabilistic estimates of climate sensitivity have also been obtained using statistical measures of the correspondence of simulated global fields to observations of time averaged present day climate (Murphy et al., 2004; Piani et al., 2005), historical transient evolution of surface temperature, upper air temperature, ocean temperature, estimates of the radiative forcing, satellite data, proxy data over the last millennium, or a subset thereof (Wigley et al., 1997a; Tol and De Vos, 1998; Andronova and Schlesinger, 2001; Forest et al., 2002; Gregory et al., 2002a; Knutti et al., 2002; Knutti et al., 2003; Frame et al., 2005; Forest et al., 2006; Forster and Gregory, 2006; Hegerl et al., 2006, see Chapter 9, Section 9.6), measures of the variability in present day climate (Bony et al., 2004; Bony and Dufresne, 2005; Williams et al., 2005), the climatological seasonal cycle of surface temperature (Tsushima et al., 2005; Knutti et al., 2006), the response to paleoclimatic forcings (Annan et al., 2005b; Hegerl et al., 2006; Schneider von Deimling et al., 2006) and major volcanic eruptions (Wigley et al., 2005; Yokohata et al., 2005, see Chapter 9, Section 9.6).

For the purpose of constraining regional climate change, spatial averages or fields of time averaged regional climate have been used (Giorgi and Mearns, 2003; Tebaldi et al., 2004; Tebaldi et al., 2006), as have regional or continental scale trends in surface temperature (Greene et al., 2006; Stott et al., 2006a). Trends in multiple variables derived from reanalysis datasets are another possibility for future consideration (Lucarini and Russell, 2002).

It is not yet clear to what extent the spread of predicted future changes can be narrowed by combining the above constraints. Additional constraints could also be found, for example from evaluation of ensemble climate prediction systems on shorter time scales for which verification data exists. These could include assessment of the reliability of seasonal to interannual probabilistic forecasts (Palmer et al., 2004; Hagedorn et al., 2005), and the evaluation of model parameterisations in short range weather predictions (Phillips et al., 2004; Palmer, 2005). There are also methodological issues to be resolved in observationally constrained model projections concerning the role and quantification of structural model errors (Goldstein and Rougier, 2005).

10.5.4.5 Probabilistic projections—global mean

A number of methods for providing probabilistic climate change projections, both for global means (discussed in this section) and geographical depictions (discussed in the following section) have emerged since the TAR. These can be grouped into three broad categories, consisting of: (1) methods based on results of AOGCM ensembles without formal application of observational constraints; (2) methods designed to be constrained by observations and their uncertainties and, as far as possible, independent of the spread of outcomes found in AOGCM ensembles; (3) methods designed to give results dependent on both observational constraints and distributions of AOGCM results.

Method 1 has the advantage of being constrained by the detailed understanding of physical processes built into the models. For example, AOGCM results from both the AR4 multi-model ensemble and from perturbed physics ensembles suggest a very low probability for a climate sensitivity below 2°C, despite exploring the effects of a wide range of alternative modelling assumptions on the global radiative feedbacks arising from lapse rate, water vapour, surface albedo and cloud (Soden and Held, 2006; Webb et al., 2006, see also Box 10.2). However, reliance on AOGCM ensembles can also be questioned on the basis that models share components, and therefore errors, and may not sample the full range of possibilities consistent with our physical understanding (e.g., Allen and Ingram, 2002). Also, the results are likely to depend on subjective judgements such as the prior distributions chosen for poorly constrained model parameters (Murphy et al., 2004; Stainforth et al., 2005).

Observationally-constrained probability distributions for climate sensitivity (method 2 above) have been derived from physical relationships based on energy balance considerations, and instrumental observations of historical changes during the past 50–150 years, or proxy reconstructions of surface temperature during the past millennium (see Chapter 9, Section 9.6). The results vary according to the choice of verifying observations, the forcings considered and their specified uncertainties, however all these studies report a high upper limit for climate sensitivity, the 95th percentile of the distributions invariably exceeding 6°C (see Box 10.2). Frame et al. (2005) demonstrate that uncertainty ranges for sensitivity are dependent on the choices made for prior distributions of uncertain quantities before the observations are applied. Frame et al. (2005) and Piani et al. (2005) show that many observable variables are likely to scale inversely with climate sensitivity, implying that projections of quantities which are inversely related to sensitivity will be more strongly constrained by observations than climate sensitivity itself, particularly with respect to the estimated upper limit (Allen et al., 2006).

In the case of transient climate change, optimal detection techniques have been used to determine factors by which hindcasts of global surface temperature from AOGCMs can be scaled up or down while remaining consistent with past changes, accounting for uncertainty due to internal variability (see Chapter 9, Section 9.4.1). Uncertainty is propagated forward in time by assuming that the fractional error found in model hindcasts of global mean temperature change will remain constant in its projections of future changes. Using this approach, Stott and Kettleborough (2002) found that probabilistic projections of global mean temperature derived from HadCM3 simulations were insensitive to differences between four representative SRES emissions scenarios over the first few decades of the 21st century; a temperature rise of 0.9–1.9°C being predicted by the 2020s relative to pre-industrial conditions. However, Stott and Kettleborough (2002) found much larger differences between the response to different SRES scenarios by the end of the 21st century (see also Section 10.5.3 and Figure 10.5.6). Stott et al. (2006b) showed that the results are relatively model-independent, finding that scaling the responses brings three models with different sensitivities into better agreement. Stott et al. (2006a) extend their approach to obtain probabilistic projections of future warming averaged over continental scale regions under the SRES A2 scenario. Fractional errors in the past continental warming simulated by HadCM3 are used to scale future changes, yielding wide uncertainty ranges, notably for North America and Europe where the 5–95% ranges for warming during the 21st century are 2–12°C and 2–11°C respectively. These estimates can be viewed as an upper limit on uncertainty as they do not account for potential constraints arising from regionally differentiated warming rates. Alternatively, a lower limit can be obtained by assuming the simulated spatial patterns of change to be correct, and scaling the regional changes according to fractional errors in global temperature change. This results in tighter ranges of 4–8°C for North America and 4–7°C for Europe.

The third method to probabilistic projection is to combine information from AOGCM ensembles and observational constraints. Allen and Ingram (2002) suggest that probabilistic projections for some variables may be made by searching for “emergent constraints”. These are relationships between variables which can be directly constrained by observations, such as global surface temperature, and variables which may be indirectly constrained by establishing a consistent, physically-based relationship which holds across a wide range of models. They present an example in which future changes in global mean precipitation are constrained using a probability distribution for global temperature obtained from a large EMIC ensemble (Forest et al., 2002) and a relationship between precipitation and temperature obtained from multi-model ensembles of the response to doubled CO_2. Further examples of this approach are those of Piani et al. (2005) and Knutti et al. (2006), described in Section 10.5.4.2. Since these methods are designed to produce distributions constrained by observations, they are likely to be robust (Allen et al., 2002; Allen and Stainforth, 2002), provided the inter-variable relationships expressing the emergent constraints are not specific to the particular ensemble of models from which they are derived.

A synthesis of published probabilistic global mean projections for the SRES scenarios B1, A1B and A2 is given in Figure 10.5.7. Probability density functions (PDFs) are given for short-term projections (2020–2030) and the end of the century (2090–2100). For comparison, normal distributions fitted to results from AOGCMs in the multi-model archive (see Section 10.3.1) are also given, though these curve fits should not be regarded as PDFs. The four methods of producing PDFs are all based on different models and/or techniques, described in Section 10.5. In short, Wigley and Raper (2001) used a large ensemble of a simple model with expert priors on climate sensitivity, ocean heat uptake, sulphate forcing and the carbon cycle, without applying constraints. Knutti et al. (2002; 2003) use a large ensemble of EMIC simulations with non-informative priors, consider uncertainties on climate sensitivity, ocean heat uptake, radiative forcing, and the carbon cycle, and apply observational constraints. Neither method considers natural variability explicitly. Stott et al. (2006b) apply the fingerprint scaling method to AOGCM simulations to obtain PDFs which implicitly account for uncertainties in forcing, climate sensitivity and internal unforced as well as forced natural variability, but neglect carbon cycle uncertainties: For the A2 scenario results obtained from three different AOGCMs are shown, illustrating the extent to which the Stott et al. PDFs depend on the model used. Harris et al. (2006) obtain PDFs by boosting a 17 member perturbed physics ensemble of the HadCM3 model using scaled equilibrium responses from a larger ensemble of simulations. This method neglects uncertainties in converting SRES emissions into forcing (e.g., from carbon cycle feedbacks). Furrer et al. (2006) use a Bayesian method described in Section 10.5.4.6 to calculate PDFs from the AR4 multi model ensemble.

[INSERT FIGURE 10.5.7 HERE]

Two key points emerge from Figure 10.5.7. For the projected near-term warming early in this century: (i) there is more agreement among models and methods (narrow width of the PDFs) compared to later in the century (wider PDFs); (ii) the near term warming early in this century is similar across different scenarios, compared to later in the century where the choice of scenario makes a real difference in the global warming that is projected to occur. These conclusions are entirely consistent with the results obtained by SCMs (Section 10.5.3).

Additionally, projection uncertainties increase close to linearly with temperature in most studies. The different methods show relatively good agreement in the shape and width of the PDFs, but with some offsets due to structural uncertainties related to the model, setup, forcing and statistics. Stott et al. (2006b) project total climate change including variations in natural forcings, resulting in small probability for cooling over the next decades. The other studies project only anthropogenic changes and exclude cooling in the future. The results of Knutti et al. (2003) show wider PDFs and probability for very high warming for the end of the century because they sample uniformly in climate sensitivities (see Chapter 9 and Box 10.2 on climate sensitivity). Resampling uniformly in observables (Frame et al., 2005) would bring those PDFs closer to the others. The other methods explicitly or implicitly exclude high sensitivities. In sum, probabilistic estimates of uncertainties for the next decades seem robust across a variety of models and methods, while results for the end of the century depend on the assumptions made. The range encompassing all PDFs for 2100 is wider than the range implied by the normal fit to the multi-model ensemble results.

10.5.4.6 Probabilistic projections—geographical depictions

Tebaldi et al. (2004) present a Bayesian approach to regional climate prediction, developed from ideas presented by Giorgi and Mearns (2002; 2003). Non-informative prior distributions for regional temperature and precipitation are updated using observations and results from AOGCM ensembles to produce probability distributions of future changes for various regions. Key assumptions are that each model and the observations differ randomly and independently from the true climate, and that the weight given to a model prediction should depend on the bias in its present day simulation and its degree of convergence with the best estimate of the predicted future change. Lopez et al. (2006) apply the Tebaldi et al. (2004) method to predict future changes in global surface temperature under a 1% per year increase in CO_2 .They compare it with the method developed by Allen et al. (2000) and Stott and Kettleborough (2002) and find that their Bayesian method (applied to a 15 member ensemble) predicts a much narrower uncertainty range than that of Allen et al. (2000) and Stott and Kettleborough (2002), which aims to provide model-independent probabilities consistent with observed changes (see above). The results of the Bayesian method are found to be sensitive to choices made in its design, in particular the convergence criterion for upweighting models close to the ensemble mean, relaxation of which substantially increases the predicted uncertainty, reducing the discrepancy with Allen et al. (2000) and Stott and Kettleborough (2002).

Another method by Furrer et al. (2006) employs a hierarchical Bayes model to construct PDFs of temperature change at each grid point from a multi-model dataset. The main assumptions are that large scales can be separated from small scales, that the true climate change signal is a common large scale structure represented to some degree in each of the model simulations, and that the signal unexplained by climate change is AOGCM specific in terms of small scale structure, but can be regarded as noise when averaged over all AOGCMs. In this method spatial fields of future minus present temperature difference from each member of a multi-model ensemble are regressed upon basis functions. One of the basis functions is a map of differences of observed temperatures from late minus mid 20th century, and others are spherical harmonics. The statistical model (formulated through a hierachical-Bayes framework, and a Markov chain Monte Carlo algorithm) then calculates estimates of the regression coefficients and their associated errors. The estimated coefficients are assumed to be centered around true (unknown) coefficients which give the true pattern of climate change as a combination of the basis functions. The error in the regression (assumed to be Gaussian) accounts for the deviation in each AOGCM from the (assumed) true pattern of change. By recombining the coefficients with the basis functions, an estimate is derived of the true climate change field and its associated uncertainty, thus providing joint probabilities for climate change at all grid points around the globe.

Estimates of uncertainty derived from multi-model ensembles of 10–20 members are potentially sensitive to outliers (Räisänen, 2001). Harris et al. (2006) propose a method of augmenting the size of AOGCM ensembles using a pattern scaling approach. They obtain frequency distributions of transient regional changes by scaling the equilibrium response patterns of a large perturbed physics ensemble. For a given model version the transient response is emulated by scaling the equilibrium response pattern according to global temperature (predicted from an energy balance model tuned to the relevant climate sensitivity), adding a correction field to account for differences between the equilibrium and transient patterns, and allowing for uncertainty in the emulated result. The correction field and emulation errors are calibrated by comparing the responses of a set of model versions (the red curves in Figure 10.5.6) for which both transient and equilibrium simulations exist. Results are used to estimate uncertainties in the surface temperature and precipitation response to increasing CO_2 arising from the combined effects of atmospheric parameter perturbations and internal variability in HadCM3.

[INSERT FIGURE 10.5.8 HERE]

This type of information can be displayed in a map of, for example, values of probability for a 2°C temperature change larger than 2°C by the end of the 21st century under the A1B scenario. Figure 10.5.8 compares probabilities estimated from the 21 member AR4 multi-model ensemble (Furrer et al., 2006) against values estimated by combining transient and equilibrium perturbed physics ensembles of 17 and 128 members respectively (Harris et al., 2006). Although the methods use different ensembles and different approaches to estimate probabilities, the large scale patterns are similar in many respects. Both methods show larger probabilities (typically 80% or more) over land, and at high latitudes in the winter hemisphere, with relatively low values (typically less than 50%) over the southern oceans. However, the plots also reveal some substantial differences at a regional level, notably over the north Atlantic ocean, the sub-tropical Atlantic and Pacific oceans in the southern hemisphere, and at high northern latitudes during June to August. See Chapter 11, Section 11.2.2.2 for further discussion of probabilistic projections of regional changes.

10.5.4.7 Summary

Significant progress has been made since the TAR in exploring ensemble approaches to provide uncertainty ranges and probabilities for global and regional climate change. Different methods show consistency in some aspects of their results, but differ significantly in others (see Box 10.2, Figures 10.5.7 and 10.5.8), because they depend to varying degrees on the nature and use of observational constraints, the nature and design of model ensembles, and the specification of prior distributions for uncertain inputs (see, for example, Chapter 11, Table 11.2.1). It is not yet possible to recommend a preferred method, but it is important to communicate to users the assumptions and limitations underlying the various approaches and the sensitivity of the results to them. A good example concerns the treatment of model error in Bayesian methods, the uncertainty in which affects the calculation of the likelihood of different model versions, but is difficult to specify (Rougier, 2006). Awareness of this issue is growing in the field of climate prediction (Annan et al., 2005b; Knutti et al., 2006), but the issue is yet to be thoroughly addressed. Probabilistic depictions, particularly at the regional level, are new to the field of climate change science and are being facilitated by the recently available multi-model ensembles. Work in this area is almost certain to develop rapidly, and will likely play a major role in how climate change is quantified and communicated in the IPCC AR5.

10.6 Sea-Level Change

10.6.1 Global Average Sea-Level Rise due to Thermal Expansion

As sea water becomes warmer, its density decreases. This thermal expansion will lead to an increase in volume of the global ocean, producing a (thermosteric) sea level rise (see Chapter 5, Section 5.5.3). Global average thermal expansion can be calculated directly from simulated changes in ocean temperature. Results are available from 16 AOGCMs for the 21st century for SRES scenarios A1B, A2 and B1 (Figure 10.6.1), continuing from simulations of the 20th century. The timeseries are rather smooth compared with global average temperature timeseries, because thermal expansion reflects heat storage in the entire ocean, being approximately proportional to the time-integral of temperature change (Gregory et al., 2001). In the control runs (available from a subset of 11 AOGCMs), internally generated variability in the climate system produces a decadal standard deviation of sea level due to thermal expansion in the range 0.4–2.0 mm, with one outlying model giving 3.4 mm.

[INSERT FIGURE 10.6.1 HERE]

During 2000–2020 under scenario SRES A1B the rate of thermal expansion is 1.4 ± 0.8 mm yr^{-1}, similar to the observationally derived rate of 1.6 ± 0.6 mm yr^{-1} during 1993–2003 (see Chapter 5, Section 5.5.3), which itself is large compared with that of previous decades, perhaps in part owing to internal variability (see Chapter 5, Section 5.5.3). Different treatment among AOGCMs of natural forcings during the 20th century (see Chapter 9, Section 9.5.2) may explain some of the spread in the early 21st century.

During 2080–2100 the rate of thermal expansion is 2.9 ± 1.9 mm yr^{-1}, more than twice that of 2000–2020, the acceleration being caused by the increased climatic warming. By 2100 thermal expansion has reached 230 ± 100 mm. There is no significant correlation of the global average temperature change across models with either thermal expansion or its rate of change, suggesting that the spread in thermal expansion is not mainly caused by the spread in surface warming, but by model-dependence in ocean heat uptake efficiency (Raper et al., 2002; Chaper 8, Table 8.2.1) and the distribution of added heat within the ocean (Russell et al., 2000).

10.6.2 Local Sea-Level Change due to Ocean Dynamics

The geographical pattern (the dynamic topography) of mean sea level relative to the geoid is an aspect of the dynamical balance relating the ocean's density structure and its circulation, which are maintained by air-sea fluxes of heat, fresh water and momentum. Over much of the ocean on multi-annual timescales, a good approximation to local sea level change is given by the steric sea level change, which can be calculated straightforwardly from local temperature and salinity change (Gregory et al., 2001; Lowe and Gregory, 2006). In much of the world, salinity changes are as important as temperature changes in determining the pattern of dynamic topography change in the future, and their contributions can be opposed (Landerer et al., 2006, and as in the past, Chapter 5, Section 5.5.4.1). Lowe and Gregory (2006) show that in the HadCM3 AOGCM changes in heat fluxes are the cause of many of the large-scale features of sea level change, but fresh water flux change dominates the North Atlantic and momentum flux change has a signature in the north Pacific and the Southern Ocean. Landerer et al. (2006) find that increased windstress in the Southern Ocean opposes thermosteric sea level change there.

Results are available for local sea level change due to ocean density and circulation change from 14 AOGCMs for the 20th century and the 21st century following SRES scenario A1B. There is substantial spatial variability in all models, i.e., sea level change is not uniform, and as the geographical pattern of climate change intensifies, the spatial standard deviation increases (Church et al., 2001; Gregory et al., 2001). Suzuki et al. (2005) show that, in their high-resolution model, enhanced eddy activity contributes to this increase. We consider the difference between the means for 2080–2100 and 1980–2000. The spatial standard deviation lies in the range 0.02–0.57 m, with an average of 0.14 m, smaller values being more common. Across models, it shows no correlation with spatial resolution but has a weak correlation of 0.4 with global average thermal expansion.

The geographical patterns of sea level change from different models are not generally similar in detail, but they have more similarity than those analysed in the TAR by Church et al. (2001). The largest spatial correlation coefficient between any pair is 0.76, but only 20% of correlation coefficients exceed 0.5. To identify common features we examine an ensemble mean (Figure 10.6.2). Like Church et al. (2001) and Gregory et al. (2001) we note smaller than average sea level rise in the Southern Ocean and larger than average in the Arctic, the former possibly due to windstress change (Landerer et al., 2006) or low thermal expansivity (Lowe and Gregory, 2006) and the latter to freshening. Another obvious feature is a narrow band of pronounced sea level rise stretching across the southern Atlantic and Indian Oceans and discernible in the southern Pacific. This could be associated with a southward shift in the circumpolar front (Suzuki et al., 2005) or subduction of warm anomalies in the region of formation of sub-Antarctic mode water (Banks et al., 2002). In the zonal mean, there are maxima of sea-level rise in 30–45°S and 30–45°N. Similar indications are present in the altimetric and thermosteric patterns of sea level change for 1993–2003 (Chapter 5, Figures 5.5.3 and 5.6.1). The model projections do not share other aspects of the observed pattern of sea level rise, such as in the western Pacific, which could be related to interannual variability.

[INSERT FIGURE 10.6.2 HERE]

The North Atlantic dipole pattern noted by Church et al. (2001), i.e., reduced rise to the south of the Gulf Stream extension, enhanced to the north, consistent with a weakening of the circulation, is present in some models; a more complex feature is described by Landerer et al. (2006). The reverse is apparent in the north Pacific, associated with a wind-driven intensification of the Kuroshio current by Suzuki et al. (2005). Using simplified models, Hsieh and Bryan (1996) and Johnson and Marshall (2002) show how upper-ocean velocities and sea level would be affected in north Atlantic coastal regions within months of a cessation of sinking in the north Atlantic as a result of propagation by coastal and equatorial Kelvin waves, but would take decades to adjust in the central regions and the south Atlantic. Levermann et al. (2005) show that a sea level rise of a several tenths of a metre could be realised in coastal regions of the North Atlantic within a few decades (i.e., tens of millimetres per year) of a collapse of the overturning. Such changes to dynamic topography would be much more rapid than global average sea level change. However, it should be emphasised that these studies are sensitivity tests, not projections; the overturning circulation does not collapse in the SRES scenario runs (see Section 10.3.4).

The geographical pattern of sea level change is affected also by changes in atmospheric pressure, land movements (elastic and viscous) resulting from the changing loading of the crust by water and ice and the consequent displacement of mantle material, and changes in the gravitational field of the ocean and solid Earth (see Chapter 5, Section 5.5.4.4). These effects have not been included in Figure 10.6.2.

10.6.3 Glaciers and Ice Caps

Glaciers and ice caps (G&IC) may change their mass because of changes in ablation (mostly melting) or accumulation (mostly precipitation), see also Chapter 4, Section 4.5.1. Since their mass balance depends strongly on their altitude and aspect, use of data from climate models to make projections requires a method of downscaling, because individual G&IC (i.e., excluding the ice sheets of Greenland and Antarctica, Section 10.6.4) are much smaller than typical AOGCM gridboxes. Statistical relations can be developed between GCM and local meteorology (Reichert et al., 2002), but they may not continue to hold in future climates. Hence for projections the approach usually adopted is to use GCM simulations of changes in climate parameters to perturb the observed climatology or mass balance (Schneeberger et al., 2003).

10.6.3.1 Mass balance sensitivity to temperature

For the TAR projections of G&IC outside Greenland and Antarctica, Church et al. (2001) used an empirical relationship (Zuo and Oerlemans, 1997) between climatological precipitation and mass balance sensitivity to temperature b_T as determined by energy-balance modelling for a sample of 12 G&IC, including refreezing of meltwater within the firn. Refer to Chapter 4, Section 4.5.1 for a discussion of the relation of b_T to climate. For a temperature change uniform throughout the year, b_T was between –0.14 and –1.00 m yr^{-1} K^{-1}. Oerlemans and Reichert (2000) and Oerlemans (2001) have refined the scheme to include dependence on monthly temperature and precipitation changes, and Oerlemans et al. (2006) have applied this version to a number of Arctic G&IC regions. Using a degree-day method (in which ablation is proportional to the integral of mean daily temperature above freezing point), de Woul and Hock (2006) find somewhat larger sensitivities for Arctic G&IC. Braithwaite et al. (2003) calculated b_T for a set of 61 G&IC using a degree-day method, with results in the range –0.1 to –1.2 m yr^{-1} K^{-1}. Braithwaite and Raper (2002) show there is excellent consistency between the results for b_T of Oerlemans and Fortuin (1992) and Braithwaite et al. (2003). Schneeberger et al. (2000; 2003) use a degree-day method for ablation modified to include incident solar radiation, again obtaining a similar range.

The sensitivity of global G&IC mass change is estimated by weighting the local sensitivities by land ice area in various regions. For a geographically and seasonally uniform rise in global temperature, Oerlemans and Fortuin (1992) derive a global G&IC mass balance sensitivity of –0.40 m yr^{-1} K^{-1}, Braithwaite and Raper (2002) –0.41, Raper and Braithwaite (2005) –0.35. Applying the scheme of Oerlemans (2001) and Oerlemans et al. (2006) worldwide gives a smaller value of –0.32 m yr^{-1} K^{-1}, the reduction being due to the improved treatment of albedo by Oerlemans (2001). Note that these sensitivities are given only for comparison of the various methods; they cannot be used for projections, which require regional and seasonal temperature changes (Gregory and Oerlemans, 1998; van de Wal and Wild, 2001).

There is considerable uncertainty in total G&IC area (see Chapter 4, Section 4.5.1 and Table 4.5.1). Assuming 510,000 km^2, a sensitivity of –0.40 m yr^{-1} K^{-1} is equivalent to 0.56 mm yr^{-1} K^{-1} of sea level rise.

Hansen and Nazarenko (2004) collated measurements of soot (fossil fuel black carbon) in snow and have estimated consequent reductions of snow and ice albedo of between 0.1% for the pristine conditions of Antarctica and over 10% for polluted northern hemisphere land areas. They argue that glacial ablation would be increased by this effect. While it is true that soot has not been explicitly considered in existing sensitivity estimates, it may already be included because the albedo and degree-day parametrisations have been empirically derived from data collected in affected regions.

10.6.3.2 Mass balance sensitivity to precipitation

For seasonally uniform temperature rise, Oerlemans et al. (1998) found that an increase in precipitation of 20–50% K^{-1} was required to balance increased ablation, while Braithwaite et al. (2003) reported of 29–41% K^{-1}, in both cases for a sample of G&IC representing a variety of climatic regimes. Oerlemans et al. (2006) require 20–43% K^{-1} and de Woul and Hock (2006) ~20% K^{-1} for Arctic G&IC. Although AOGCMs generally indicate larger than average precipitation in northern mid- and high-latitude regions, the global average is less than 5% K^{-1}, so we would expect ablation increases to dominate worldwide. However, precipitation changes may locally be important (see Chapter 4, Section 4.5.3).

10.6.3.3 Dynamic response

As glacier volume is lost, glacier area declines so the ablation decreases. Oerlemans et al. (1998) calculated that omitting this effect leads to overestimates of ablation of about 25% by 2100. Church et al. (2001), following Bahr et al. (1997) and Van de Wal and Wild (2001), made some allowance for it by diminishing the area A of a glacier of volume V according to $V \propto A^{1.375}$. This is a scaling relation derived for glaciers in a steady state, which may hold approximately during retreat, but comparison with a simple flow model suggests the deviations do not exceed 20% (van de Wal and Wild, 2001). Schneeberger et al. (2003) found that the scaling relation produced a mixture of over- and under-estimates of volume loss for their sample of glaciers by comparison with more detailed modelling.

A further serious difficulty is that b_T of the glacier should change as volume is lost: lowering the ice surface as the glacier thins will tend to make b_T more negative, but the predominant loss of area at lower altitude in the ablation zone will tend to make it less negative (Braithwaite and Raper, 2002). The latter effect is more important (Schneeberger et al., 2003) for two reasons: lowering the surface at the high rate of 1 m yr^{-1}, for instance, will raise its temperature by <0.01 K yr^{-1}, which is small compared with the projected rate of climatic warming, and those areas of the ablation zone which thin most rapidly will soon be removed altogether. Raper et al. (2000) applied a further scaling relation to reduce the width and length of a glacier as its area declines, while keeping its maximum altitude and shape fixed. This treatment includes the main influences on b_T, but still excludes the time-dependent effects of glacier dynamics.

The geometrical and dynamical approaches of Raper et al. (2000) and Oerlemans et al. (1998) cannot be applied to all the world's glaciers individually as the required data are unknown for the vast majority of them. Instead, it might be applied to a representative ensemble derived from statistics of size distributions of G&IC. A geometrical calculation in which glaciers are treated statistically and ice-caps individually shows that the reduction of area of glaciers strongly reduces the ablation during the 21st century (Raper and Braithwaite, 2006). They obtain 46 mm and 51 mm of sea-level rise under scenario SRES A1B for the GFDL-CM2.0 and PCM AOGCMs. Similar results of 55 mm for both are given by the mass-balance sensitivities to temperature of Oerlemans (2001) and Oerlemans et al. (2006) with the area-scaling of Van de Wal and Wild (2001) and the G&IC volume of Raper and Braithwaite (2005) (see Chapter 4, Table 4.5.1), while the results without area-scaling are 85 and 87 mm. This comparison suggests that the area-scaling and the geometrical model have a similar effect in reducing estimated ablation for the 21st century.

10.6.3.4 Global average sea-level change due to glacier and ice-cap changes

Applying the mass balance sensitivities of Oerlemans (2001) and Oerlemans et al. (2006) worldwide with the area-scaling of Van de Wal and Wild (2001) to temperature and precipitation changes simulated by 17 AOGCMs under scenarios A1B, A2 and B1 gives a global G&IC mass balance sensitivity b_g to global temperature change of 0.36 ± 0.14 mm yr^{-1} K^{-1}. It is 0.49 ± 0.16 mm yr^{-1} K^{-1} excluding precipitation change, which in this scheme offsets 2550% of the melting. This is larger than expected from idealised studies (see

Section 10.6.3.2) because precipitation increases are projected to be larger than average in glaciated regions. Regressing observed global G&IC mass balance changes against global temperature change up to the mid-1990s suggests a somewhat larger value of 0.65 ± 0.40 mm yr^{-1} K^{-1} (see Chapter 9, Section 9.5.2), although the model b_g is statistically consistent with this, given the large observation uncertainties. The increase of G&IC mass loss in more recent years would support a higher value of ~0.8 mm yr^{-1} K^{-1}. Contributions to this may come from local decadal climatic variations, and in some regions where glaciers discharge into the sea or lakes there is accelerated dynamic discharge (Rignot et al., 2003) that is not included in currently available glacier models. Given these uncertainties, and considering also for the ±30% uncertainty estimated for the mass balance calculation (Gregory and Oerlemans, 1998; Raper and Braithwaite, 2006), we adopt b_g=0.65 ± 0.30 mm yr^{-1} K^{-1} for projections for the G&IC outside Greenland and Antarctica.

Using a degree-day scheme, Vaughan (2006a) estimates that ablation of glaciers in the Antarctic Peninsula presently amounts to 0.055 mm yr^{-1} of sea level. However, Morris and Mulvaney (2004) find that accumulation increases have been larger than ablation increases during 1972–1998, giving a small net negative sea level contribution from the region. Because ablation increases non-linearly with temperature, for future warming the contribution would become positive; they estimate a contribution of 0.07 ± 0.02 mm yr^{-1} K^{-1} to sea level rise, i.e., about 10% of the aggregate of other G&IC. A model estimate for the G&IC separate from the ice sheet on Greenland indicates a further addition of about 6% to the sea level contribution in the 21st century (van de Wal and Wild, 2001). In recent decades, the G&IC on Greenland and Antarctica have together added about 20% (see Chapter 4, Section 4.5.2). On these grounds, we increase the G&IC sea-level contribution by a factor of 1.2 to allow for them.

We make G&IC projections assuming that the climate of 1901–1930 was 0.15 K warmer than the steady-state for glaciers (c.f., Zuo and Oerlemans, 1997; Gregory et al., 2006). Although results for the 20th century are rather sensitive to this assumption, the projected temperature change in the 21st century is large by comparison. We use three alternative estimates of world glacier volume (see Chapter 4, Table 4.5.1) for the area-scaling. Under scenario SRES A1B, the projected contribution to sea level change during the 21st century is 88 ± 68 mm. Lower values are projected than by Church et al. (2001) because the more recent estimates of total glacier volume are smaller, implying a more rapid wastage of area.

10.6.4 Ice Sheets

The mass of ice grounded on land in the Greenland and Antarctic ice sheets could change as a result of changes in surface mass balance (SMB, the sum of accumulation and ablation) or in the flux of ice crossing the grounding line, which is determined by the dynamics of the ice sheet (see Section 10.6.5). Surface mass balance and dynamics together determine, and are both affected by, the change in surface topography.

10.6.4.1 Surface mass balance

SMB is immediately influenced by climate change. A good simulation of the ice sheet SMB requires a resolution exceeding that of AGCMs used for long climate experiments, because of the steep slopes at the margins of the ice sheet, where the majority of the precipitation and all of the ablation occurs. Orographic forcing of precipitation is typically overestimated by AGCMs, whose smooth topography does not present a sufficient barrier to inland penetration (Ohmura et al., 1996b; Glover, 1999; Murphy et al., 2002). Ablation also tends to be overestimated because the area at low altitude around the margins of the ice sheet is exaggerated, where melting preferentially occurs (Glover, 1999; Wild et al., 2003). In addition, AGCMs do not generally have a representation of the refreezing of surface meltwater within the snowpack and may not include albedo variations dependent on snow ageing and its conversion to ice.

To address these issues, several groups have computed SMB at resolutions of tens of kilometres or less, with results that compare acceptably well with observations (e.g., van Lipzig et al., 2002; Wild et al., 2003). Ablation is calculated either by schemes based on temperature (degree-day or other temperature-index methods), calibrated from observations (Braithwaite, 1995; Ohmura et al., 1996a), or by energy-balance modelling (stand-alone or within a GCM). Changes in SMB have been studied using climate change simulated by high-resolution AGCMs or by perturbing an observational climatology with lower-resolution climate model output.

There is considerable uncertainty in accumulation changes (Table 10.6.1, van de Wal et al., 2001; Huybrechts et al., 2004). Precipitation increase could be determined by atmospheric radiative balance, increase in saturation specific humidity with temperature, circulation changes, retreat of sea ice permitting greater evaporation, or a combination of these (van Lipzig et al., 2002). Accumulation also depends on change in local temperature, which strongly affects whether it falls as solid or liquid (Janssens and Huybrechts, 2000). For Greenland, accumulation derived from the high-resolution AGCMs increases by 5–9% K^{-1}. Precipitation increases somewhat less in AR4 AOGCMs (typically of lower resolution), by 4–7% K^{-1}. Kapsner et al. (1995) do not find a relationship between precipitation and temperature variability inferred from Greenland ice cores for the Holocene, although both show large changes from the LGM to the Holocene. In the HadCM3 AOGCM relationship is strong for climate change forced by greenhouse gases and the glacial-interglacial transition, but weaker for naturally forced variability (Gregory et al., 2006). For Antarctica, precipitation increases at 6–9% K^{-1} in the time-slices and 3–8% K^{-1} in the AOGCMs.

Since there will not be substantial ablation in Antarctica, all studies for the 21st century find that Antarctic SMB changes contribute negatively to sea level, owing to increasing accumulation (see Chapter 4, Section 4.6.3 for comparison with changes in the last decade). However, in Greenland, ablation is important. Table 10.6.1 shows uncertainty in the ablation, which is particularly sensitive to temperature change around the margins. Climate models give smaller warming in these low-altitude regions than for the Greenland average, and smaller warming in summer (when ablation occurs) than on the annual average, but larger warming in Greenland than on the global average (Church et al., 2001; Huybrechts et al., 2004; Chylek and Lohmann, 2005; Gregory and Huybrechts, 2006). In most studies Greenland SMB changes give a net positive contribution to sea level in the 21st century, because the ablation increase is larger than the precipitation increase. Only Wild et al. (2003) find the opposite, so that the net SMB change contributes negatively to sea level in the 21st century. Wild et al. (2003) attribute this difference to the reduced ablation area on their higher-resolution grid. A positive SMB change is not consistent with analyses of recent changes in Greenland SMB (see Chapter 4, Section 4.6.3).

Table 10.6.1. Comparison of ice sheet (grounded ice area) surface mass balance changes calculated from high-resolution climate models. $\Delta P/\Delta T$ is the change in accumulation divided by change in temperature over the ice sheet, expressed as sea level equivalent (positive for falling sea level), and $\Delta R/\Delta T$ the corresponding quantity for ablation (positive for rising sea level). Note that ablation increases more rapidly than linearly with ΔT (van de Wal et al., 2001; Gregory and Huybrechts, 2006). To convert to kg yr^{-1} K^{-1}, multiply by 3.6×10^{14} m^2. To convert to mm yr^{-1} K^{-1} averaged over the ice sheet, multiply by 206 for Greenland and 26 for Antarctica. $\Delta P/P\Delta T$ is the fractional change in accumulation divided by the change in temperature.
[a]In these cases P is precipitation rather than accumulation.

Study	Climate model	SMB from energy balance or temperature index	Greenland $\Delta P/\Delta T$ (mm yr^{-1} K^{-1})	Greenland $\Delta P/P\Delta T$ (% K^{-1})	Greenland $\Delta R/\Delta T$ (mm yr^{-1} K^{-1})	Antarctica $\Delta P/\Delta T$ (mm yr^{-1} K^{-1})	Antarctica $\Delta P/P\Delta T$ (% K^{-1})
Van de Wal et al. (2001)	ECHAM4	20 km EB	0.14	8.5	0.16	–	–
Wild and Ohmura (2000)	ECHAM4	T106 ≈1.1° EB	0.13	8.2	0.22	0.47	7.4
Wild et al. (2003)	ECHAM4	2 km TI			0.04		
Bugnion and Stone (2002)	ECHAM4	20 km EB	0.10	6.4	0.13	–	–
Huybrechts et al. (2004)	ECHAM4	20 km TI	0.13[a]	7.6[a]	0.14	0.49[a]	7.3[a]
Huybrechts et al. (2004)	HadAM3H	20 km TI	0.09[a]	4.7[a]	0.23	0.37[a]	5.5[a]
Van Lipzig et al. (2002)	RACMO	55 km EB	–	–	–	0.53	9.0

10.6.4.2 Dynamics

Ice-sheet flow reacts to changes in topography produced by SMB change. Using a thermomechanical ice sheet model, Huybrechts et al. (2004) found that increased discharge from Antarctica during the 21st century offset a small part (<5%) of the increased accumulation, while dynamical change reduced by 10–20% the mass loss from Greenland (Huybrechts et al., 2002). Such dynamical responses are relatively minor. The TAR concluded that accelerated sea level rise caused by rapid dynamic response of the ice sheets to climate change is very unlikely during the 21st century (Church et al., 2001). However, new evidence of rapid recent changes in the Antarctic Peninsula, West Antarctica and Greenland (see Chapter 4, Section 4.6) has raised again the possibility of larger dynamical changes in future centuries (Vaughan, 2006b) than are projected by state-of-the-art continental models, such as used by Huybrechts et al. (2004), because these simulations do not incorporate all the processes responsible for the rapid marginal thinning currently taking place.

Regarding ice dynamics, the main uncertainty is the degree to which the presence of ice shelves affects the flow of inland ice across the grounding line. Theoretical studies suggest that the ice shelves can exist without transmitting back-stress to ice sheets (e.g., Hindmarsh, 1993). Comparisons between measurements and models of streaming and non-streaming flows (Whillans and van der Veen, 1993; Mayer and Huybrechts, 1999) also indicate that back-stress can be insignificant. However, a strong argument for enhanced flow when the ice shelf is removed is yielded by the acceleration of Jakobshavn Glacier (Greenland) following the loss of its floating tongue, and of the glaciers supplying the Larsen-B ice shelf (Antarctic Peninsula) after it collapsed (see Chapter 4, Section 4.6.3.3). The onset of disintegration of the Larsen-B ice shelf has been attributed to enhanced fracturing by crevasses promoted by surface meltwater (Scambos et al., 2000). Large portions of the Ross and Filchner ice shelves currently have mean summer surface temperatures of around –5°C, with the Ronne ~2°C colder (Comiso, 2000, updated). High-resolution GCMs (Gregory and Huybrechts, 2006) indicate that summer surface warming in these regions is 60 ± 40% of the Antarctica annual average, which in turn is a factor 1.1 ± 0.4 greater than global average warming in AOGCM simulations under SRES scenarios to 2100 followed by stabilisation to 2300. The uncertainties are not well characterised but these figures suggest that a local summer warming of 5°C is very unlikely for a global warming of less than 5°C. Under the A2 scenario, for instance, it is therefore unlikely that during the 21st century these ice shelves will experience frequent surface melting that might induce distintegration. In the Amundsen Sea sector of West Antarctica, ice-shelves are not so extensive and the cause of ice-shelf thinning is not surface melting, but bottom melting at the grounding line (Rignot and Jacobs, 2002). Shepherd et al. (2004) give an average ice-shelf thinning rate of 1.5 ± 0.5 m yr^{-1}.At the same time as the basal melting, accelerated inland flow has been observed for Pine Island, Thwaites and other glaciers in the sector (Rignot, 1998, 2001; Thomas et al., 2004). The synchroneity of these changes strongly implies that their cause lies in the area's ice shelves caused by oceanographic change in the Amundsen Sea, but this has not been attributed to anthropogenic climate change and could be connected with variability in the Southern Annular Mode.

Because the acceleration took place in only a few years (Rignot et al., 2002; Joughin et al., 2003) but appears up to ~150 km inland, it implies that the dynamical response to changes in the ice shelf can propagate rapidly up the ice stream. This conclusion is supported by modelling studies of Pine Island Glacier by Payne et al. (2004) and Dupont and Alley (2005), in which basal or lateral drag at the ice front is instantaneously reduced in idealised ways. The simulated acceleration and inland thinning are rapid but transient; the rate of contribution to sea level declines as a new steady state is reached over a few decades. In the study of Payne et al. (2004) the imposed perturbations were designed to resemble loss of drag in the "ice plain", a partially grounded region near the ice front, and produced a velocity increase of ~1 km yr^{-1} there; Thomas et al. (2005) suggest the ice plain will become ungrounded during the next decade and obtain a similar velocity increase using a simplified approach.

Most of inland ice of West Antarctica is grounded below sea level so could be floated if it thinned sufficiently; discharge therefore promotes inland retreat of the grounding line, which represents a positive feedback by further reducing basal traction (although it does not itself alter sea level). Unlike the one-off instantaneous change in the idealised studies, this would represent a sustained dynamical forcing that would prolong the contribution to sea-level rise. Grounding-line retreat of the ice streams has been observed recently at up to ~1 km yr^{-1} (Rignot, 1998, 2001; Shepherd et al., 2002) but is difficult to model (Vieli and Payne, 2005).

The observation in west-central Greenland of seasonal variation in ice flow rate and of a correlation with summer temperature variation (Zwally et al., 2002) has prompted speculation that surface meltwater is able to penetrate more than 1200 m of subfreezing ice to join a subglacially routed drainage system lubricating the ice flow. However, other studies (Echelmeyer and Harrison, 1990; Joughin et al., 2004) found no evidence for seasonal fluctuations in the flow rate of nearby Jakobshavn Glacier despite a substantial supply of surface meltwater; thus other explanations for flow variability cannot be discounted.

10.6.4.3 Global average sea-level rise due to ice-sheet changes

Computing ice-sheet surface mass balance using four high-resolution AGCM simulations with the scaling technique of Huybrechts et al. (2004), see Section 10.6.4.1, indicates systematic uncertainties of ±60% for Greenland and ±40% for Antarctica associated with the geographical and seasonal pattern of climate change over the ice sheets (derived from results of Gregory and Huybrechts, 2006). From 18 AR4 AOGCMs, the mass balance sensitivities are 0.18 ± 0.14 mm yr^{-1} K^{-1} for Greenland and -0.31 ± 0.24 mm yr^{-1} K^{-1} for Antarctica, evaluated for an average temperature change of 3°C over each ice-sheet. These results generally cover the range shown in Table 10.6.1, but tend to give more positive (Greenland) or less negative (Antarctica) sea level rise because of the smaller precipitation increases predicted by the AOGCMs than in the high-resolution AGCMs. We include a further uncertainty of 20% of the Greenland ablation to allow for uncertainty in the parametrisation (Church et al., 2001).

Dynamic changes simulated by a continental ice sheet model (Huybrechts and De Wolde, 1999) can be roughly represented as modifying the sea-level changes due to SMB change by $-40 \pm 20\%$ m^{-1} from Antarctica, and ±20% from Greenland (Gregory and Huybrechts, 2006, see Section 10.6.4.2), for contributions of up to ~0.5 m. Including these effects, the sea level contributions during the 21st century under the SRES A1B scenario are 35 ± 32 mm from Greenland and -55 ± 52 mm from Antarctica. During 2080–2100 their rates of contribution are 0.7 ± 0.4 mm yr^{-1} and -0.9 ± 0.7 mm yr^{-1} respectively.

The net ice sheet contribution is thus likely to be relatively small, unless larger dynamic changes occur of the kind currently observed in some Greenland outlet glaciers and West Antarctic ice streams (see Section 10.6.3.2). A sensitivity study for Greenland (Parizek and Alley, 2004) suggested that acceleration of ice flow due to meltwater reaching the bed might increase its sea-level contribution during the 21st century by up to 0.2 m. For Antarctica, quantitative uncertainties are such that we can only indicate orders of magnitude. The majority of West Antarctic ice discharge is through the ice streams which feed the Ross and Ronne-Filchner ice shelves, but in these regions no accelerated flow causing thinning is currently observed; on the contrary, they are thickening or near balance (Zwally et al., 2005). Excluding these regions, and likewise those parts of the East Antarctic ice sheet which drain into the large Amery ice shelf, the total area of ice streams (areas flowing faster than 100 m yr^{-1}) discharging directly into the sea or via a small ice shelf is 270,000 km^2. If all these areas thinned at 2 m yr^{-1}, the order of magnitude of the larger rates observed in fast-flowing areas of the Amundsen Sea sector (Shepherd et al., 2001; Shepherd et al., 2002), the contribution to sea level rise would be ~1.5 mm yr^{-1}. This would require sustained retreat simultaneously on many fronts, and should be taken as an indicative upper limit for the 21st century, not a projection; it is ~10 times greater than the rate of loss of inland ice during the 1990s from accelerated flow, which was 0.13 ± 0.02 mm yr^{-1} (Shepherd et al., 2002). A lower limit is given by assuming that the present dynamical imbalance is a response to variability, rather than a trend, and will diminish during coming decades. The state of understanding prevents a best estimate from being made, but we note that the dynamical imbalance would have to increase by a factor of ~5 to outweigh the mid-range projection under scenario A1B for increased accumulation.

10.6.5 Projections of Global Average Sea-Level Change for the 21st Century

Combining the results for thermal expansion, glacier and ice sheet mass change obtained in Sections 10.6.1, 10.6.3 and 10.6.4 gives projections under the medium emission scenarios (SRES A1B) of global average sea-level rise with respect to 2000 of 34 ± 25 mm by 2020, 120 ± 60 mm by 2050, and 290 ± 150 mm by 2100, at a rate of 3.6 ± 2.1 mm yr^{-1} during 2080–2100. This rate is considerably above the 1961–2003 average rate of 1.8 ± 0.5 mm yr^{-1} (see Chapter 5, Section 5.5.2.1). If we use only those models whose results for thermal expansion and glacier melt during 1961–2003 fall within the observational ranges (see Chapter 5, Section 5.5.6), we obtain 320 ± 110 mm by 2100, and 4.1 ± 1.2 mm yr^{-1} during 2080–2100. Furthermore, the sum of the terms for 1961–2003 both for observations (see Chapter 5, Section 5.5.6) and models (see Chapter 9, Section 9.5.2) is ~0.7 mm yr^{-1} less than the observed rate of sea-level rise. The discrepancy is statistically consistent with zero, but if it really arises from an unaccounted term in the budget which persists into the future, it implies an extra ~70 mm by 2100. The 1993–2003 rate of 3.1 ± 0.8 mm yr^{-1} (see Chapter 5, Section 5.5.2.2) may have a substantial contribution from internally generated decadal variability (see Chapter 5, Section 5.5.2.4 and Chapter 9, Section 9.5.2). If the difference between the 1961–2003 and 1993–2003 rates is caused entirely by variability rather than forced climate change, it implies that decadal rates of sea-level rise may differ from multidecadal trends by ~1 mm yr^{-1}.

For an average model, the scenario spread (from minimum to maximum) in sea level rise is only 8 mm at 2020. This is small because of the time-integrating effect of sea level rise, on which the divergence among the scenarios has had little effect by then. At 2050 it is 50 mm and at 2100 it is 200 mm.

Thermal expansion is the largest component. Accelerated flow in the Greenland and Antarctic ice-sheets could increase their contributions substantially but quantitative projections cannot be made with confidence (see Section 10.6.4.3). The ongoing response of the ice sheets to paleoclimate change would add 10 ± 10 mm by 2100 (see Chapter 6, Section 6.4.3). Thawing of permafrost is projected to contribute about 5 mm during the 21st century under scenario SRES A2 (calculated from Lawrence and Slater, 2005). The mass of the ocean will also be changed by climatically driven alteration in other water storage, in the forms of atmospheric water vapour, seasonal snow cover, soil moisture, groundwater, lakes and rivers. All of these are expected to be relatively small terms, but there may be substantial contributions from anthropogenic change in terrestrial water storage, through extraction from aquifers and impounding in reservoirs (see Chapter 5, Section 5.5.5.3).

10.7 Long Term Climate Change and Commitment

10.7.1 Climate Change Commitment Out to Year 2300 Based on AOGCMs

Building on Wigley (2005) we use three specific definitions of climate change commitment: (i) the "constant composition commitment" which denotes the further change of temperature (*constant composition temperature commitment*), sea level (*constant composition sea level commitment*), or any other quantity in the climate system, since the time the composition of the atmosphere, and hence the radiative forcing, has been held at a constant value; (ii) the "constant emission commitment" which denotes the further change of, e.g., temperature (*constant emission temperature commitment*) since the time the greenhouse gas emissions have been held at a constant value; and (iii), the "zero emission commitment" which denotes the further change of, e.g., temperature (*zero emission temperature commitment*) since the time the greenhouse gas emissions have been set to zero.

The concept that the climate system exhibits commitment when radiative forcing has changed, is mainly due to the thermal inertia of the oceans, and was discussed independently by Wigley (1984), Hansen et al. (1984), and Siegenthaler and Oeschger (1984). The term “commitment” in this regard was introduced by Ramanathan (1988). In the TAR this was illustrated in idealized scenarios of doubling and quadrupling CO_2, and stabilization at 2050 and 2100 after an IS92a forcing scenario. Various temperature commitment values were reported (about 0.3°C per century with much model-dependency), and EMIC simulations were used to illustrate long-term influence of the ocean owing to long mixing times and meridional overturning circulation. Subsequent studies have confirmed this behavior of the climate system and ascribed it to the inherent property of the climate system that the thermal inertia of the ocean introduces a lag to the warming of the climate system after concentrations of greenhouse gases are stabilized (Mitchell et al., 2000; Wetherald et al., 2001; Wigley and Raper, 2003; Hansen et al., 2005b; Meehl et al., 2005c; Wigley, 2005). Climate change commitment as discussed here should not be confused with "unavoidable climate change" over the next half century, which would surely be greater because forcing cannot be instantly stabilized. Furthermore, in the very long term it is plausible that climate change could be less than in a commitment run since forcing could plausibly be reduced below current levels (i.e., see WG2, Chapter 2, Section 2.3.1.2) as illustrated in the overshoot simulations and zero emission commitment simuations discussed below.

Three constant composion commitment experiments have recently been performed by the global coupled climate modeling community: (1) stabilizing concentrations of GHGs at year 2000 values after a 20th century climate simulation, and running an additional 100 years; (2) stabilizing concentrations of GHGs at year 2100 values after a 21st century B1 experiment and running an additional 100 years (with some models run to 200 years); and (3) stabilizing concentrations of GHGs at year 2100 values after a 21st century A1B experiment, and running an additional 100 years (and some models to 200 years). Multi-model mean warming in these experiments is depicted in Figure 10.3.2. Time series of the globally averaged surface temperature and percent precipitation change after stabilization are shown for all the models in Figure 10.7.1. The multi-model average warming in the first experiment reported earlier for several of the models (Meehl et al., 2005c) is about 0.5°C at year 2100, nearly the magnitude of warming simulated in the 20th century.

Hansen et al. (2005a) calculate the current energy imbalance of the Earth to be 0.85 W m^{-2}, implying that the unrealized global warming is about 0.6°C without any further increase in radiative forcing.

[INSERT FIGURE 10.7.1 HERE]

For the B1 commitment run, the additional warming after 100 years is also about 0.5°C, and roughly the same for the A1B commitment (Figure 10.7.1). These new results quantify what was postulated in the TAR in that warming commitment after stabilizing concentrations is about 0.5°C for the first century, and considerably smaller after that, with most of the warming commitment occurring in the first several decades of the 22nd century.

Constant composition precipitation commitment for the multi-model ensemble average is about 1.1% by 2100 for the 20th century commitment experiment, and for the B1 commitment experiment by 2200 is 0.8% and by 2300 is 1.5%, while for the A1B constant composition commitment experiment by 2200 is 1.5% and 2% by 2300.

The patterns of change in temperature in the B1 and A1B experiments, relative to pre-industrial, do not change greatly after stabilization, as is quantified in Table 10.3.2. Even the 20th century stabilization case warms with some similarity to the A1B pattern (Table 10.3.2). However, there is some contrast in the land and ocean warming rates, as seen from Figure 10.3.3. Mid and low latitude land warms at rates closer to the global mean of that of A1B, while high latitude ocean warming is larger.

10.7.2 Climate Change Commitment Out to Year 3000 and Beyond to Equilibrium

EMICs are used to extend the projections for a scenario that follows A1B to 2100 and then keeps atmospheric composition, and hence radiative forcing, constant out to the year 3000. By 2100 the projected warming is between 1.2 and 4.1°C, similar to the range projected by AOGCMs. A large constant composition temperature and sea level commitment is evident in the simulations and is slowly realized over coming centuries. By the year 3000 the warming range is 1.9 to 5.6°C. While surface temperatures approach equilibrium relatively quickly, sea level continues to rise for many centuries.

Five of these EMICs include interactive representations of the marine and terrestrial carbon cycle and, therefore, can be used to assess carbon cycle-climate feedbacks and effects of carbon emission reductions on atmospheric CO_2 and climate. Although carbon cycle processes in these models are simplified, global-scale quantities are in good agreement with more complex models (Doney et al., 2004).

[INSERT FIGURE 10.7.2 HERE]

Results for one carbon emission scenario are shown in Figure 10.7.3 where anthropogenic emissions follow a path towards stabilization of atmospheric CO_2 at 750 ppm but at year 2100 are reduced to zero. This permits the determination of the zero emission climate change commitment. The prescribed emissions were calculated from the SP750 profile (Knutti et al., 2005) using the Bern Carbon Cycle Model (Joos et al., 2001). Although unrealistic, such a scenario permits the calculation of zero emission commitment, i.e., climate change due to 21st century emissions. Even though emissions are instantly reduced to zero at year 2100, it takes about 100 to 400 years in the different models for the atmospheric CO_2 concentration to drop from the maximum (ranges between 650 to 700 ppm) to below the level of two times preindustrial CO_2 (~560 ppm) owing to a continuous transfer of carbon from the atmosphere into the terrestrial and oceanic reservoirs. Emissions effected in the 21st century continue to have an impact even at year 3000 when both surface temperature and sea level rise due to thermal expansion are still substantially higher than preindustrial. Also shown are atmospheric CO_2 concentrations and ocean/terrestrial carbon inventories at year 3000 versus total emitted carbon for similar emission pathways targeting 450, 550, 750 and 1000 ppm atmospheric CO_2 and with carbon emissions reduced to zero at year 2100. Atmospheric CO_2 at year 3000 is approximately linearly related to the total amount of carbon emitted in each model, but with a substantial spread among the models in both slope and absolute values, because the redistribution of carbon between the different reservoirs is model dependent. In summary, the model results show that 21st century emissions represent a minimum commitment of climate change for several centuries, irrespective of later emissions. A reduction of this "minimum" commitment is possible only if, in addition to avoiding CO_2 emissions after 2100, CO_2 were actively removed from the atmosphere.

[INSERT FIGURE 10.7.3 HERE]

Using a similar approach, Friedlingstein and Solomon (2005) showed that even if emissions were immediately cut to zero, the system would continue to warm for several more decades before starting to cool. It is important also to note that ocean heat content and changes in the cryosphere evolve on time scales extending over centuries.

On very long timescales (order four thousand years as estimated by an AOGCM experiment), equilibrium climate sensitivity is a useful concept to characterize the ultimate response of climate models to different future levels of greenhouse gas radiative forcing. This concept can be applied to climate models irrespective of their complexity. Based on a global energy balance argument, equilibrium climate sensitivity S and global mean surface temperature increase ΔT at equilibrium relative to preindustrial for an equivalent stable CO_2 concentration are linearly related according to $\Delta T = S \times \log(CO_2/280\ \text{ppm})/\log(2)$. This concept can be used to illustrate the temperature change expected at equilibrium and assuming that the linearity of the relationship holds true. Because the combination of various lines of modelling results and expert judgement yields a quantified range of climate sensitivity S (see Box 10.2), this can be carried over to equilibrium temperature increase. Most likely values, and the likely range, as well as a very likely lower bound for the warming, all consistent with the quantified range of S, are given in Table 10.7.1.

Table 10.7.1. Best guess, likely and very likely bounds/ranges of global mean equilibrium surface temperature increase ΔT over preindustrial temperatures for different levels of CO_2 equivalent radiative forcing. Numbers are generated from the equation $\Delta T = S \times \log(CO_2/280\ \text{ppm})/\log(2)$, where S is the equilibrium climate sensitivity, and using the expert assessment that S is most likely around 3°C, likely between 2 and 4.5°C, and very likely above 1.5°C (see Box 10.2). In contrast to the projections for SRES and other emissions scenarios, these numbers are not based directly on climate model simulations, but follow from the assumption that the concept of climate sensitivity from which the above equation is derived, and the estimates of its value provided in this report are correct.

Eq CO_2	**best guess**	**very likely above**	**likely between**	**and**
350	1.0	0.5	0.6	1.4
450	2.1	1.0	1.4	3.1
550	2.9	1.5	1.9	4.4
650	3.6	1.8	2.4	5.5
750	4.3	2.1	2.8	6.4
1000	5.5	2.8	3.7	8.3
1200	6.3	3.1	4.2	9.4

It is emphasized that this table does not contain more information than our best knowledge of S and that the numbers are not the result of any climate model simulation. Rather it is assumed that the simple relationship $\Delta T = S \times \log(CO_2/280\ \text{ppm})/\log(2)$ holds true for the entire range of equivalent CO_2 concentrations. There are limitations to the concept of radiative forcing and climate sensitivity (Senior and Mitchell, 2000; Joshi et al., 2003; Shine et al., 2003; Hansen et al., 2005b). Only a few AOGCMs have been run to equilibrium under elevated CO_2 concentrations, and some results show that nonlinearities in the feedbacks (e.g., clouds, sea ice and snow cover) may cause a time dependence of the effective climate sensitivity and substantial deviations from the linear relation assumed above (Manabe and Stouffer, 1994; Senior and Mitchell, 2000; Voss and Mikolajewicz, 2001; Gregory et al., 2004b), with effective climate sensitivity tending to grow with time in many of the AR4 AOGCMs. Some studies suggest that climate sensitivities larger than the likely estimate given below (which would suggest greater warming) cannot be ruled out (see Box 10.2 on climate sensitivity).

Another way to address eventual equilibrium for different CO_2 concentrations is to use the projections from the AOGCMs in Figure 10.3.1, forced with an idealized 1% per year CO_2 increase to 4 × CO_2. The equivalent CO_2 concentrations in the AOGCMs can be estimated from the forcings given in Table 6.14 in the TAR. The actual CO_2 concentrations for A1B and B1 are roughly 715 ppm and 550 ppm (depending on which model is used to convert emissions to concentrations), and equivalent CO_2 concentrations are estimated to be about 778 ppm and 561 ppm, respectively. Using the equation above for an equilibrium climate sensitivity of 3.0°C, eventual equilibrium warming in these experiments would be 4.4°C and 3.0°C, respectively. The multi-model average warming in the AOGCMs at the end of the 21st century (Table 10.3.2) is 2.6°C and 1.8°C, or close to 60% of the eventual estimated equilibrium warming. Given rates of CO_2 increase of between 0.5% and 1.0% in these two scenarios, this can be compared to the calculated fraction of eventual warming of around 50% in AOGCM experiments with those CO_2 increase rates (Stouffer and Manabe, 1999). That model had somewhat higher equilibrium climate sensitivity, and was actually run to equilibrium in a 4000 year integration to enable comparison of transient and equilibrium warming. Therefore, the AOGCM results combined with the estimated equilibrium warming seem roughly consistent with earlier AOGCM experiments of transient warming rates. Additionally, we can compute similar numbers for the 4×CO_2 stabilization experiments performed with the AOGCMs. In that case the actual and equivalent CO_2 concentrations are the same, since there are no other radiatively active species changing in the models, and the multi-model CO_2 concentration at quadrupling would produce an eventual equilibrium warming of 6°C, where the multi-model average warming at the time of quadrupling is about 4.0°C or 66% of eventual equilibrium. This is roughly consistent with the numbers for the A1B and B1 scenario integrations with the AOGCMs.

So now we can estimate how much closer to equilibrium the climate system is 100 years after stabilization in these AOGCM experiments. After 100 years of stabilized concentrations, the warming has increased to 3.3°C in A1B or nearly 75% of the estimated eventual equilibrium warming, and in B1 the warming is 2.1°C or about 70% of the estimated equilibrium warming. For the stabilized 4×CO_2 experiment, after 100 years of stabilized CO_2 concentrations the warming is 4.7°C, or 78% of the estimated equilibrium warming. Therefore, about an additional 10% of the eventual equilibrium warming is achieved after 100 years of stabilized concentrations. This emphasizes that the approach to equilibrium takes a long time, and even after 100 years of stabilized concentrations, only about 70% to 80% of the eventual equilibrium warming is realized.

10.7.3 Long-Term Integrations: Idealized Overshoot Experiments

The concept of mitigation related to overshoot scenarios has implications for WG2 and WG3 and was addressed already in the SAR. A new suite of mitigation scenarios is currently being assessed for the AR4. WG1 does not have the expertise to assess such scenarios, so here we assess the processes and response of the physical climate system in a very idealized overshoot experiment. Plausible new mitigation and overshoot scenarios subsequently will be run by modelling groups in WG1 and assessed in the next IPCC report.

An idealized overshoot scenario has been run in an AOGCM where the concentrations reduce from the A1B stabilized level to the B1 stabilized level between 2150 and 2250 followed by 200 years of integration with that constant B1 level of concentrations (Figure 10.7.4a). This reduction in concentrations would require huge and likely implausible reductions in emissions, but such an idealized experiment illustrates the processes involved with how the climate system would respond to such a large change in emissions and concentrations. Yoshida et al. (2005) show there is a relatively fast response in the surface and upper ocean in starting to recover to temperatures at the B1 level after several decades, but a much more sluggish response with more commitment in the deep ocean. As shown in Figure 10.7.4b and c, the overshoot scenario temperatures only slowly reduce to approach the lower temperatures of the B1 experiment, and continue a slow convergence that has still not cooled to the B1 level at the year 2350, or 100 years after the CO_2 concentrations in the overshoot experiment were reduced to equal the concentrations in the B1 experiment. However, Dai et al. (2001b) have shown that reducing emissions to achieve a stabilized level of concentrations in the 21st century reduces warming moderately (less than 0.5°C) by the end of the 21st century in comparison to a business-as-usual scenario, but the warming reduction is about 1.5°C by the end of the 22nd century in that experiment. Other climate system responses include the North Atlantic MOC and sea ice volume that almost recover to the B1 level in the overshoot scenario experiment, except for a significant hysteresis effect that is shown in the sea level change due to thermal expansion (Yoshida et al., 2005)

[INSERT FIGURE 10.7.4 HERE]

Such stabilization and overshoot scenarios have implications for risk assessment as suggested by Yoshida et al. (2005) and others. For example, in a probabilistic study using an SCM and multi-gas scenarios, Meinshausen (2006) estimated that the probability of overshooting a 2°C warming is between 68% and 99% for a stabilization of equivalent CO_2 at 550 ppm. They also considered scenarios with peaking CO_2 and subsequent stabilization at lower levels as an alternative pathway and found that if the risk of exceeding a warming of 2°C is not to exceed 30%, it is necessary to peak equivalent CO_2 concentrations around 475 ppm before returning to lower concentrations of about 400 ppm. These overshoot and targeted climate change estimations take into account the climate change commitment in the system that must be overcome on the timescale of any overshoot or emissions target calculation. The probabilistic studies also show that when certain thresholds of climate change are to be avoided, emission pathways depend on the certainty requested of not overshooting the threshold.

Intermediate complexity models (EMICs) have been used to calculate the long-term climate response to stabilization of atmospheric CO_2, though EMICs have not been adjusted to take into account the full range of AOGCM sensitivities. The newly developed stabilization profiles were constructed following Enting et al. (1994) and Wigley et al. (1996) using the most recent atmospheric CO_2 observations, CO_2 projections with the Bern Carbon Cycle-Climate model (Joos et al., 2001) for the A1T scenario over the next few decades, and a ratio of two polynomials (Enting et al., 1994) leading to stabilization at levels of 450, 550, 650, 750 and 1000 ppm atmospheric CO_2 equivalent. Other forcings are not considered. Figure 10.7.5a shows the equilibrium surface warming for seven different EMICs and six stabilization levels. Model differences arise mainly from the models having different climate sensitivities.

Knutti et al. (2005) explored this further in an EMIC using several published PDFs of climate sensitivity and different ocean heat uptake parameterizations and calculated probabilities of not overshooting a certain temperature threshold given an equivalent CO_2 stabilization level (Figure 10.7.5b). This plot illustrates, for example, that for low values of stabilized CO_2, the range of response of possible warming is smaller than for high values of stabilized CO_2. This is because with greater CO_2 forcing, there is a greater spread of outcomes as was illustrated in Figure 10.5.2. It also shows that for any given temperature threshold, the smaller the probability of overshooting the target should be, the lower the stabilization level must be chosen.

[INSERT FIGURE 10.7.5 HERE]

10.7.4 Commitment to Sea-Level Change

10.7.4.1 Thermal expansion

The sea level rise commitment due to thermal expansion has much longer timescales than the warming commitment, owing to the slow processes which mix heat into the deep ocean (Church et al., 2001). Despite stabilisation of concentrations, thermal expansion in the 22nd and 23rd centuries is greater than in the 21st century in most models, as noted by Meehl et al (2005c), reaching 0.3–0.8 m by 2300 in A1B (Figure 10.7.6). There is a wide spread among the models for the thermal expansion commitment due partly to climate sensitivity, partly to differences in the parameterization of vertical mixing affecting ocean heat uptake, as shown by Weaver and Wiebe (1999) for instance. If there is deep water formation in the final steady state as in the present day, the ocean will eventually warm up fairly uniformly by the amount of the global average surface temperature change (Stouffer and Manabe, 2003), which would give about 0.5 m K^{-1} of thermal expansion calculated from observed climatology; the EMICs in Figure 10.7.2 indicate 0.3-0.6 m K^{-1}. If deep water formation is weakened or suppressed, the deep ocean will warm up more (Knutti and Stocker, 2000). For instance, in the $3 \times CO_2$ experiment of Bi et al. (2001) with the CSIRO AOGCM, both NADW and AABW formation cease, and the steady-state thermosteric sea level rise is 4.5 m.

[INSERT FIG 10.7.6 HERE]

10.7.4.2 Glaciers and ice caps

Steady-state projections for G&IC require a model which evolves their area-altitude distribution, since even in a warmer climate, some glacier volume may persist at high altitude. Little information is available on this. A recent study (Raper and Braithwaite, 2006) found that with a geographically uniform warming relative to 1900 of 4°C maintained after 2100, ~60% of G&IC volume would vanish by 2200 and practically all by 3000.

10.7.4.3 Greenland ice sheet

The present SMB of Greenland is a net accumulation estimated as 0.6 mm yr^{-1} of sea level equivalent from a compilation of studies (Church et al., 2001) and 0.47 mm yr^{-1} for 1988–2004 (Box et al., 2006). In a steady state the net accumulation would be balanced by calving of icebergs. GCMs suggest that accumulation increases linearly with temperature (van de Wal et al., 2001; Gregory and Huybrechts, 2006), whereas ablation increases more rapidly, so warming will tend to reduce the SMB, as has been observed in recent years (see Chapter 4, Section 4.6.3), and sufficient warming will reverse its sign. The warming that reduces the SMB to zero gives a threshold for the long-term viability of the ice sheet, because negative SMB means that the ice sheet must contract even if ice discharge has ceased owing to retreat from the coast. If a warmer climate is maintained, the ice sheet will eventually be eliminated, except perhaps for remnant glaciers in the mountains, raising sea-level by ~7 m (see Chapter 4, Table 4.1.1). Huybrechts et al. (1991) evaluated the threshold as 2.7°C in seasonally and geographically uniform warming over Greenland. Gregory et al. (2004a) examined the probability of this threshold being reached under various CO_2 stabilisation scenarios for 450–1000 ppm using TAR projections, finding that it was passed for 34 out of 35 combinations of AOGCM and CO_2 concentration considering seasonally uniform warming, and 24 out of 35 considering summer warming and using an upper bound on the threshold.

Because warming is predicted to be weaker in the ablation area and in summer (see Section 10.6.4.1), using high-resolution geographical and seasonal patterns of simulated temperature change, rather than assuming it to be uniform, raises the threshold to 4.5 ± 1.8°C in annual- and area-average warming in Greenland, and 3.1 ± 1.6°C in the global average (Gregory and Huybrechts, 2006), relative to pre-industrial (assumed to be a steady state) i.e., 2.3 ± 1.6°C in the global average relative to today (see Chapter 3, Section 3.2.2.6). This is likely to be reached by 2100 under scenario SRES A1B, for instance (Figure 10.5.3). These results are supported by evidence from the last interglacial, when the temperature in Greenland was 2–4°C warmer than today and the ice sheet survived, but was smaller by 2.2–3.5 m of sea level (including contributions from Arctic ice caps, see Chapter 6, Section 6.4.3). However, a lower threshold of 1°C (Hansen, 2005) in global warming has also been suggested, on the basis that global (rather than Greenland) temperatures during previous interglacials exceeded today's by no more than that.

For a global warming of 3°C relative to present, models suggest Greenland would contribute 0.2–3.9 mm yr^{-1} to sea level. The greater the warming, the faster the loss of mass. Ablation would be further enhanced by the lowering of the surface, which is not included in the calculations. To include this and other climate feedbacks in calculating long-term rates of sea level rise requires coupling an ice-sheet model to a climate model. Ridley et al. (2005) coupled the Greenland ice sheet model of Huybrechts and De Wolde (1999) to the HadCM3 AOGCM Under constant 4 × CO_2, the sea level contribution was 5.5 mm yr^{-1} over the first 300 years and declined as the ice sheet contracted; after 1000 years only about 40% of the original volume remained and after 3000 years only 4% (Figure 10.7.7). The rate of deglaciation would be increased if ice-flow was accelerated by basal lubrication due to surface meltwater (see Section 10.6.4.2). The best estimate of Parizek and Alley (2004) was that this could add an extra 0.15–0.40 m to sea level by 2500, compared with 0.4–3.2 m calculated by Huybrechts and De Wolde (1999) without this effect. The processes whereby meltwater might penetrate through subfreezing ice to the bed are unclear and only conceptual models exist at present (Alley et al., 2005).

[INSERT FIGURE 10.7.7 HERE]

Even with pre-industrial or present-day CO_2, the climate of Greenland would be much warmer without the ice sheet, because of lower surface altitude and albedo, and there is medium likelihood that Greenland deglaciation and the resulting sea level rise would be irreversible even if global climate change were reversed. Toniazzo et al. (2004) found that snow does not accumulate anywhere on the ice-free Greenland with pre-industrial CO_2, whereas Lunt et al. (2004) obtained a substantial regenerated ice sheet in east and central Greenland, using a higher-resolution model.

10.7.4.4 Antarctic ice sheet

GCMs indicate increasingly positive SMB for the Antarctic ice sheet as a whole with rising global temperature, because of greater accumulation, contributing up to several mm yr^{-1} of sea level fall (1.2 ± 1.1 mm yr^{-1} for a global warming of 3°C relative to present) as computed by Gregory and Huybrechts (2006). Continental ice-sheet models indicate this would be offset by tens of percent by inreased ice discharge, but still giving a negative contribution to sea level, of –0.8 m by 3000 in one simulation with Antarctic warming of ~4.5°C (Huybrechts and De Wolde, 1999).

However, discharge could increase substantially if buttressing due to the major West Antarctic ice shelves were reduced (see Section 10.6.4.2), and could outweigh the accumulation increase, leading to a net positive Antarctic sea-level contribution on the long term. If the Amundsen Sea sector were eventually deglaciated, it would add ~1.5 m to sea level, while the entire WAIS would account for ~5 m (Vaughan, 2006b). Contributions could also come in this manner from the limited marine-based portions of East Antarctica that discharge into large ice-shelves.

Weakening or collapse of the ice shelves could be caused by surface melting or thinning due to basal melting. In equilibrium experiments with mixed-layer ocean models, the ratio of Antarctic to global annual warming is 1.4 ± 0.4. With summer warming over the major West Antarctic ice shelves being 60 ± 40% of the Antarctic annual average (see Section 10.6.4.2), it appears that mean summer temperatures in these regions are unlikely to reach melting point while global warming is less than 5°C. Comparison of several ice shelves in the Amundsen Sea Sector and modelling studies indicate that basal melt rates depend on water temperature near to the base, with a constant of proportionality of ~10 m yr^{-1} K^{-1} (Rignot and Jacobs, 2002; Shepherd et al., 2004). If this order of magnitude applies to future changes, a warming of ~1°C under the major ice shelves would eliminate them within centuries. We not able to relate this quantitatively to global warming with any confidence, because the issue has so far received little attention, and current models may be inadequate to treat it, because of limited resolution and poorly understood processes. Nonetheless it is reasonable to suppose that sustained global warming would eventually lead to warming in the sea water circulating beneath the ice shelves.

In the absence of appropriate dynamical models, there is little agreement about what dynamical changes could occur (cf. Vaughan and Spouge, 2002). One line of argument is to consider an analogy with palaeoclimate. Global average sea level stood 4–6 m higher during the last interglacial; since this cannot all be explained by Greenland, the Antarctic ice sheet must also have been smaller (see Chapter 6, Section 6.4.3). On this basis, using the limited available evidence, sustained global warming of 2°C (Oppenheimer and Alley, 2005) has been suggested as a threshold beyond which there will be a commitment to large sea-level contribution from the WAIS. Since the maximum rates of sea level rise during previous glacial terminations were of the order of magnitude of 10 mm yr^{-1} (see Chapter 6, Section 6.4.3), and since no observed recent acceleration has exceeded a factor of ten, we can be confident that future accelerated discharge from WAIS will not exceed this size, which is roughly an order of magnitude increase in present-day WAIS discharge.

Another line of argument is that there is no evidence that rates of dynamical discharge of this magnitude can be sustained over long periods. The West Antarctic ice-sheet is 20 times smaller than the LGM northern hemisphere ice sheets which contributed most of the meltwater during previous deglaciations, whose rates can be explained by surface melting alone (Zweck and Huybrechts, 2005). In the study of Huybrechts and De Wolde (1999), the largest rate of sea-level rise was 2.5 mm yr^{-1}. This was dominated by dynamical discharge associated with grounding-line retreat (see Section 10.6.4.2). The model did not simulate ice-streams, whose widespread acceleration could possibly give larger rates, but the maximum loss of ice possible from rapid discharge of existing ice streams is the volume in excess of flotation in the regions occupied by these ice streams (defined as regions of flow exceeding 100 m yr^{-1}, see Section 10.6.4.2). This volume (in both West and East Antarctica) is 230,000 km^3, equivalent to ~0.6 m of sea level, or ~1% of the mass of the Antarctic ice sheet, most of which does not flow in ice streams. Rapid loss of ice could be sustained beyond this level only if new ice streams developed in currently slow-moving ice, but the possible extent and rate of such changes cannot presently be estimated, since there is only very limited understanding of controls on the development and variability of ice streams. On this argument, rapid discharge is likely to be transient and the long-term sign of the Antarctic contribution to sea level depends on whether increased accumulation is more important than large-scale retreat of the grounding line.

Box 10.1: Future Abrupt Climate Change, "Climate Surprises", and Irreversible Changes

Theoretical studies and model simulations, as well as the abundant evidence of high-resolution proxy data from numerous paleoclimatic archives (see Chapter 6) have established the fact that changes in the climate system can be abrupt and widespread. A working definition of "abrupt climate change" was given in Alley et al. (2002): “Technically, an abrupt climate change occurs when the climate system is forced to cross some threshold, triggering a transition to a new state at a rate determined by the climate system itself and faster than the cause”. More generally, a gradual change in some determining quantity of the climate system (e.g., radiation balance, land surface properties, sea ice, etc.) can cause a variety of structurally different responses (Box 10.1, Figure 1). The response of a purely linear system scales with the forcing, and at stabilisation of the forcing, a new equilibrium is achieved which is structurally similar, but not necessarily close to the original state. However, if the system contains more than one stable equilibrium state, transitions to structurally different states are possible. Upon the crossing of a bifurcation point the evolution of the system is no longer controlled by the time scale of the forcing, but rather determined by its internal dynamics, which can either be much faster than the forcing, or significantly slower. Only the former case would be termed "abrupt climate change", but the latter case is of equal importance. For the long-term evolution of a climate variable one must distinguish between reversible and irreversible changes. Irreversibility is only a theoretical concept and assumes no additional changes in climate forcing on millennial time scales and longer. The notion "climate surprises" usually refers to abrupt transitions and temporary or permanent transitions to a different state in parts of the climate system.

[INSERT BOX 10.1, FIGURE1 HERE]

Atlantic meridional overturning circulation and other ocean circulation changes:
The best documented type of abrupt climate change in the paleoclimatic archives is that associated with changes in the ocean circulation. The climate system has thus demonstrated, that such changes were physically possible in the past (Stocker, 2000). Since TAR many new results from climate models of different complexity have provided a more detailed view on the anticipated changes of the Atlantic meridional overturning circulation (MOC) in response to global warming. Most models agree that the MOC weakens over the next 100 years (Figure 10.3.13). This weakening evolves on the time scale of the warming and ranges from indistinguishable from natural variability to about 60% by 2100. None of the AOGCM simulations shows an abrupt change when forced with the SRES emissions scenarios until 2100, but some long-term model simulations suggest that a complete cessation can result for large forcings (Stouffer and Manabe, 2003). Models of intermediate complexity have been used to explore parameter space more completely and indicate that thresholds in MOC may be present but that they depend on the amount and rate of warming for a given model. More importantly, the existence of such thresholds crucially depends on model parameterisations, e.g., the amount of vertical and horizontal mixing that is simulated in the ocean model components (Manabe and Stouffer, 1999; Knutti et al., 2000; Longworth et al., 2005). The few long-term simulations of AOGCMs indicate that even complete shutdowns of the MOC may be reversible (Stouffer and Manabe, 2003; Yoshida et al., 2005; Stouffer et al., 2006). However, until millenial simulations with AOGCMs are available, the important question of potential irreversibility of an MOC shutdown remains unanswered. Both simplified models and AOGCMs agree, however, that a potential spin-down of the MOC, induced by global warming, would take many decades to more than a century to fully spin down. There is no direct model evidence that the MOC could collapse within a few decades in response to global warming. However, a few studies do show the potential for rapid changes in the MOC (Manabe and Stouffer, 1999), and the processes concerned are poorly understood (see Chapter 8, Section 8.7). This is not inconsistent with the paleoclimate records. The cooling events during the last ice ages registered in the Greenland ice cores developed over a couple of centuries to millennia. In contrast, there were also a number of very rapid warmings, the so called Dansgaard-Oeschger events (NorthGRIP Members, 2004), which evolved on decades or less, most probably associated with rapid switch-ons of the MOC (see Chapter 6, Section 6.3.2).

Recent simulations with models, whose ocean components resolve topography in sufficient detail, obtain a consistent pattern of a strong to complete reduction of convection in the Labrador Sea (Wood et al., 1999; Schweckendiek and Willebrand, 2005). Such changes in the convection, with implications to the atmospheric circulation, can develop within a few years (Schaeffer et al., 2002). The long-term and regional-to-hemispheric scale effects of such changes in water mass properties have not yet been investigated.

With a reduction of the MOC, the meridional heat flux also reduces in the subtropical and mid latitudes with large-scale effects on the atmospheric circulation. In consequence, the warming of the North Atlantic surface proceeds more slowly. Even for strong reductions in MOC towards the end of the 21st century, no cooling is observed in the regions around the North Atlantic because it is overcompensated by the radiative forcing that caused the ocean response in the first place. In the high latitudes, an increase in the oceanic meridional heat flux is simulated by these models. This increase is due to both an increase in the overturning circulation in the Arctic and the advection of warmer waters from lower latitudes and thus contributes significantly to continuing sea ice reduction in the Atlantic sector of the Arctic (Hu et al., 2004a). Few simulations have also addressed the changes to overturning in the South Atlantic and Southern Ocean. In addition to water mass modifications, this also has an effect on the transport by the Antarctic Circumpolar Current, but results are not yet conclusive.

Arctic sea ice:
Arctic sea ice is responding sensitively to global warming. While changes in winter sea ice cover are moderate, late summer sea ice is projected to disappear almost completely towards the end of the 21st century. A number of positive feedbacks in the climate system accelerate the melt back of sea ice. The ice albedo feedback allows open water to receive more heat from the sun during summer, and the increase of ocean heat transport to the Arctic through the advection of warmer waters and stronger circulation further reduce ice cover. Minimum Arctic sea ice cover is observed in September. Model simulations indicate that the September sea ice cover reduces substantially in response to global warming. The reduction generally evolves on the time scale of the warming. With sustained warming, the late summer disappearance of a major fraction of Arctic sea ice is permanent.

Glaciers and ice caps:
Glaciers and ice caps are sensitive to changes in temperature and precipitation. Observations point to a reduction in volume over the last 20 years (see Chapter 4, Section 4.5.2), with a rate during 1993–2003 (corresponding to (0.81 ± 0.43) mm/yr sea level), with a larger mean central estimate than that for 1961–1998 (corresponding to (0.51 ± 0.32) mm/yr sea level). Rapid changes are therefore already under way and enhanced by positive feedbacks associated with the surface energy balance of shrinking glaciers and newly exposed land surface in periglacial areas. Acceleration of glacier loss over the next few decades is likely (see Section 10.6.3). Based on simulations of 11 glaciers in various regions, a volume loss of 60% of these glaciers is projected by the year 2050 (Schneeberger et al., 2003). Glaciated areas in the Americas are also affected. A comparative study including 7 GCM simulations at $2 \times CO_2$ conditions inferred that many glaciers may disappear completely due to an increase of the equilibrium line altitude (Bradley et al., 2004). The disappearance of these ice bodies is much faster than a potential reglaciation several centuries hence, and may, in some areas actually be irreversible.

Greenland and West Antarctic Ice Sheets:
Satellite and in situ measurement networks have demonstrated increasing melting and accelerated ice flow around the periphery of the Greenland Ice Sheet (GIS) over the past 25 years (see Chapter 4, Section 4.6.2). The few simulations of long-term ice sheet simulations suggest that the Greenland Ice Sheet (GIS) will significantly reduce in volume and area over the coming centuries if warming is sustained (Gregory et al., 2004a; Huybrechts et al., 2004; Ridley et al., 2005). A threshold of annual mean warming of (4.5 ± 1.8) °C in Greenland was estimated for elimination of the GIS (Gregory and Huybrechts, 2006, see Section 10.7.3.3). The melting would not proceed abruptly but take many centuries to complete. Even if temperatures were to decrease later, the reduction of the GIS to a much smaller extent might be irreversible, because the climate of an ice-free Greenland could be too warm for accumulation; however, this result is model-dependent (see Section 10.7.3.3). The positive feedbacks involved here are that once the ice sheet gets thinner, temperatures in the accumulation region are higher, increasing the melting and causing more precipitation to fall as rain rather than snow, that the lower albedo of the exposed ice-free land causes a local climatic warming; and that surface meltwater might accelerate ice flow (see Section 10.6.4.2).

A collapse of the West Antarctic Ice Sheet (WAIS) has been discussed as a potential response to global warming for many years (Bindschadler, 1998; Oppenheimer, 1998; Vaughan, 2006b). A complete disintegration of the WAIS would cause a global sea level rise of about 5 meters. The observed acceleration of ice streams in the Amundsen Sea sector of the WAIS, the rapidity of propagation of this signal upstream, and the acceleration of glaciers which fed the Larsen-B ice shelf after its collapse have renewed these concerns (see Section 10.6.4.2). It is possible that the presence of ice shelves tends to stabilize the ice sheet, at least regionally. Therefore, through the creation of surface meltwater ponds and bottom melting by a warmer ocean, the disappearance of large ice shelves might contribute to a potential destabilization of the WAIS, which could proceed through the positive feedback of grounding-line retreat. Present understanding is insufficient for prediction of the possible speed or extent of such a collapse (see Section 10.7.3.4).

Vegetation cover:
Irreversible and relatively rapid changes in vegetation cover and composition have occurred frequently in the past. The most prominent example is the desertification of the Sahara region about 5000 years ago (Claussen et al., 1999). The reason for this behaviour is believed to lie in the limitation of plant communities with respect to temperature and precipitation. Once critical levels are crossed, certain species can no longer compete within their ecosystem. Areas close to vegetation boundaries will experience particularly large and rapid changes due to the slow migration of these boundaries induced by global warming. A climate model simulation into the future shows that drying and warming in South America leads to a continuous reduction in the forest of Amazonia (Cox et al., 2000; Cox et al., 2004). While evolving continuously over the 21st century, such a change and ultimate disappearance could be irreversible, though this result could be model-dependent since analysis of 11 AOGCMs show a wide range of future possible rainfall changes over the Amazon (Li et al., 2006).

One of the possible "climate surprises" concerns the role of the soil in the global carbon cycle. As the concentration of CO_2 is increasing, the soil is acting, in the global mean, as a carbon sink by assimilating carbon due to accelerated growing of the terrestrial biosphere. However, by about 2050, a model simulation suggests that the soil changes to a source of carbon by releasing previously accumulated carbon due to increased respiration (Cox et al., 2000), induced by increasing temperature and precipitation. This represents a positive feedback to the increase in atmospheric CO_2. While different models agree regarding the sign of the feedback, large uncertainties exist regarding the strength (Cox et al., 2000; Friedlingstein et al., 2001; Dufresne et al., 2002). However, the respiration increase is caused by warmer and wetter climate. The switch from moderate sink to strong source of atmospheric carbon is rather rapid and occurs within two decades (Cox et al., 2004), but the timing of the onset is uncertain (Huntingford et al., 2004). A model intercomparison reveals that once set in motion, the increase in respiration continues even after the CO_2 levels are held constant (Cramer et al., 2001). Although considerable uncertainties still exist, it is clear that feedback mechanisms between the terrestrial biosphere and the physical climate system exist, which can qualitatively and quantitatively alter the response to an increase in radiative forcing.

Atmospheric and ocean-atmosphere regimes:
Changes in weather patterns and regimes can be abrupt processes which might occur spontaneously due to dynamical interactions in the atmosphere-ice-ocean system, or they manifest the crossing of a threshold in the system due to slow external forcing. Such shifts have been reported in SST in the tropical Pacific leading into a phase of more ENSO (Trenberth, 1990), or in the stratospheric polar vortex (Christiansen, 2003), a shut-down of deep convection in the Greenland Sea (Bönisch et al., 1997; Ronski and Budeus, 2005) and an abrupt freshening of the Labrador Sea (Dickson et al., 2002). The freshening evolves in the entire depth but the shift in salinity was particularly rapid: the 34.87 isohaline plunges from seasonally surface to 1600 meters within 2 years with no return since 1973.

In a long, unforced model simulation, a period of a few decades with anomalously cold temperatures (up to 10 standard deviations below average) in the region south of Greenland was found (Hall and Stouffer, 2001). It was caused by persistent winds which changed the stratification of the ocean and inhibited convection thereby reducing heat transfer from the ocean to the atmosphere. Similar results were found in a different model in which the major convection site in the North Atlantic spontaneously switched to a more southerly location for several decades to centuries (Goosse et al., 2002). Other simulations show that the slowly increasing radiative forcing is able to cause transitions in the convective activity in the GIN Sea which has an influence on the atmospheric circulation over Greenland and western Europe (Schaeffer et al., 2002). The changes unfold within a few years and indicate that the system has crossed a threshold.

A multi-model analysis of regimes of polar variability (NAO, AO, and AAO) reveals that the simulated trends in the 21st century are significant for the AO and AAO and point towards more zonal circulation (Rauthe et al., 2004). Temperature changes associated with changes in atmospheric circulation regimes such as NAO can easily exceed in certain regions (e.g., Northern Europe) the long-term global warming which cause such interdecadal regime shifts (Dorn et al., 2003).

Box 10.2: Climate Sensitivity

The range for equilibrium climate sensitivity was estimated in the TAR (Cubasch et al., 2001) to be 1.5 to 4.5°C, with a 1 in 3 probability of values outside that range. The range was nearly the same as an early report of the National Research Council (Charney, 1979), and the two previous IPCC assessment reports (Mitchell et al., 1990; Kattenberg et al., 1996). These estimates were expert assessments largely based on equilibrium climate sensitivities simulated by atmospheric GCMs coupled to non-dynamic slab oceans. The mean plus-minus one standard deviation of the values from these models was (3.8 ± 0.78) °C in the SAR (17 models), (3.5 ± 0.92) °C in the TAR (15 models) and now amounts to (3.2 ± 0.72) °C in 18 models.

Considerable work has been done since the TAR (2001b) to estimate climate sensitivity and to provide a better quantification of relative probabilities, including a most likely value, rather than just a subjective range of uncertainty. Since climate sensitivity of the real climate system cannot be measured directly, new methods have been used since the TAR (2001b) to establish a relationship between sensitivity and some observable quantity (either directly or through a model), and to estimate a range or probability density function (PDF) of climate sensitivity consistent with observations. These methods are summarized separately in Chapters 9 and 10, and here we synthesize that information into an assessment. The information comes from two main categories: constraints from past climate change on various timescales, and the spread of results for climate sensitivity from ensembles of models.

The first category of methods (see Chapter 9, Section 9.6) uses the historical transient evolution of surface temperature, upper air temperature, ocean temperature, estimates of the radiative forcing, satellite data, proxy data over the last millennium, or a subset thereof to calculate ranges or PDFs for sensitivity (e.g., Wigley et al., 1997b; Tol and De Vos, 1998; Andronova and Schlesinger, 2001; Forest et al., 2002; Gregory et al., 2002a; Knutti et al., 2002; Knutti et al., 2003; Frame et al., 2005; Forest et al., 2006; Forster and Gregory, 2006; Hegerl et al., 2006). A summary of all PDFs of climate sensitivity from those methods is shown in Chapter 9, Figure 9.6.1 and in Box 10.2, Figure 1a. Median values, most likely values (modes) and 5–95% confidence ranges are shown in Box 10.2, Figure 1b for each PDF. Most of the results confirm previous judgements that climate sensitivity is very unlikely below 1.5°C. The upper bound is more difficult to constrain because of a nonlinear relationship between climate sensitivity and the observed transient response, and is further hampered by limited length of the observational record and uncertainties in the observations, which are particularly large for ocean heat uptake and for the magnitude of the aerosol radiative forcing. Studies that take all the important known uncertainties in observed historical trends into account cannot rule out the possibility that the climate sensitivity exceeds 4.5 °C, though such high values are consistently found to be less likely than values of around 2.0 to 3.5°C. Observations of transient climate change provide better constraints for the transient climate response (see Chapter 9, Section 9.6.1.3)

[INSERT BOX 10.2, FIGURE 1 HERE]

Two recent studies use a modelled relation between climate sensitivity and tropical sea surface temperatures (SST) in the Last Glacial Maximum (LGM) and proxy records of the latter to estimate ranges of climate sensitivity (Annan et al., 2005b, see Chapter 9, Section 9.6; Schneider von Deimling et al., 2006). While both of these estimates overlap with results from the instrumental period and results from other AOGCMS, the results differ substantially due to the different relationships between LGM SSTs and sensitivity in the models used. Therefore, LGM proxy data provide support for the range of climate sensitivity based on other lines of evidence, but do not constrain it further.

Studies comparing the observed transient response of surface temperature after large volcanic eruptions with results obtained from models with different climate sensitivities (see Chapter 9, Section. 9.6) do not provide PDFs, but find best agreement with sensitivities around 3°C, and reasonable agreement within the 1.5–4.5°C range (Wigley et al., 2005). They are not able to exclude sensitivities above 4.5°C.

The second category of methods examines climate sensitivity in GCMs. Climate sensitivity is not a single tuneable parameter in these models, but depends on many processes and feedbacks. Three PDFs of climate sensitivity were obtained by comparing different variables of the simulated present-day climatology and variability against observations in a perturbed physics ensemble (Murphy et al., 2004; Piani et al., 2005; Knutti et al., 2006, Box 10.2, Figure 1c/d, see Section 10.5.4.2). Equilibrium climate sensitivity is found to be most likely around 3.2°C, and very unlikely below about 2°C. The upper bound is sensitive to how model parameters are sampled and to the method used to compare with observations.

Box 10.2, Figure1e and f show the frequency distributions obtained by different methods when perturbing parameters in the HadAM3 model but before weighting with observations (Chapter 10). Murphy et al. (2004)(unweighted) sampled 29 parameters and assumed individual effects to combine linearly. Stainforth et al. (2005) found nonlinearities when simulating multiple combinations of a subset of key parameters. The most frequently occurring climate sensitivity values are grouped around 3°C, but this could reflect the sensitivity of the unperturbed model. Some but not all of the high-sensitivity models have been found to agree poorly with observations and are therefore unlikely, hence even very high values are not excluded. This inability to rule out very high values is common to many methods, since for well understood physical reasons, the rate of change (against sensitivity) of most quantities that we can observe tends to zero as the sensitivity increases (Hansen et al., 1985; Knutti et al., 2005; Allen et al., 2006).

There is no well-established formal way of estimating a single PDF from the individual results, taking account of the different assumptions in each study. Most studies do not account for structural uncertainty, and thus probably tend to underestimate the uncertainty. On the other hand, since several largely independent lines of evidence indicate similar most likely values and ranges, climate sensitivity values are likely to be better constrained than those found by methods based on single datasets (Annan and Hargreaves, 2006; Hegerl et al., 2006).

The equilibrium climate sensitivity values for the AR4 GCMs coupled to non-dynamic slab ocean models are given for comparison (Box 10.2, Figure 1e and f, see also Table 8.8.1). These estimates come from models that represent the current best efforts from the international global climate modelling community at simulating climate. A normal and log-normal fit yield 5–95% ranges of about 2 to 4.5°C (Räisänen, 2005b), with median values of equilibrium climate sensitivity of about 3.2°C. A probabilistic interpretation of the results is problematic, because each model is assumed to be equally credible and the results depend upon the assumed shape of the fitted distribution. Although the AOGCMs used in IPCC reports are an ‘ensemble of opportunity’ not designed to sample modelling uncertainties systematically or randomly, the range of sensitivities covered has been rather stable over many years. This occurs in spite of substantial model developments, considerable progress in simulating many aspects of the large-scale climate, and evaluation of those models against observations. The spread of model estimates of climate sensitivity arises mostly from inter-model differences in the shortwave component of cloud radiative feedbacks, primarily due to inter-model differences in the response of low-level clouds (see Chapter 8, Section 8.6)

Since the TAR, our level of scientific understanding and confidence in quantitative estimates of equilibrium climate sensitivity have increased substantially. Based on the several independent lines of evidence, as summarized in Box 10.2, Figures 1 and 2, we conclude:

An expert assessment based on the available constraints from observations and the strength of known feedbacks simulated in GCMs indicates that the equilibrium warming for doubling carbon dioxide, or "climate sensitivity", is likely to lie in the range 2 to 4.5°C, with a most likely value of about 3°C. Climate sensitivity is very unlikely to be less than 1.5°C. For fundamental physical reasons as well as data limitations, values substantially higher than 4.5°C still cannot be excluded, but agreement with observations and proxy data is generally worse for those high values than for values in the 2 to 4.5°C range.

[INSERT BOX 10.2, FIGURE 2 HERE

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Question 10.1: Are Extreme Events, Like Heat Waves, Droughts, or Floods, Expected to Change as the Earth's Climate Changes?

Yes, extreme events are expected to change as Earth's climate changes, and these changes could occur even with relatively small mean climate changes.

In a future warmer climate, most AOGCMs simulate summer dryness and winter wetness in most parts of northern middle and high latitudes. Summer dryness indicates a greater risk of drought. Going along with the risk of drying is also an increased chance of intense precipitation and flooding. Though somewhat counter-intuitive, this is because precipitation is concentrated into more intense events, with longer periods of little precipitation in between. Therefore, intense and heavy episodic rainfall events are interspersed with longer relatively dry periods. Another aspect of these projected changes has been related to the mean changes of precipitation, with wet extremes projected to become more severe in many areas where mean precipitation is expected to increase, and dry extremes where the mean precipitation is projected to decrease.

Going along with the results for increased extremes of intense precipitation, even if the storms in future climate did not change much in intensity, there would be an increase in extreme rainfall intensity with the extra-tropical storms, particularly over Northern Hemisphere land, with an increase in the likelihood of very wet winters over much of central and northern Europe due to the increase of intense precipitation. These events would be associated with midlatitude storms, suggesting an increased chance of flooding over Europe and other midlatitude regions due to more intense rainfall and snowfall events producing more runoff. Similar results apply for summer precipitation with implications for more flooding in the Asian monsoon region and other tropical areas. The increased risk of floods in a number of major river basins in a future warmer climate has been related to an increase in river discharge with an enhanced risk for future intense storm-related precipitation events and flooding.

In a future climate there is an increased risk of more intense, longer-lasting and more frequent heat waves. The European 2003 heat wave has been used as an example of the type of extreme heat event lasting from several days to over a week that is likely to become more common in a future warmer climate. A related aspect of temperature extremes is that there is likely to be a decrease in diurnal temperature range in most regions in a future warmer climate. It is also likely that a future warmer climate would be characterized by a decrease in frost days (e.g., nights where the temperature dips below freezing). A quantity related to frost days is growing season length, and this has been projected to increase in future climate. Additionally, there is likely to be a decline in frequency of cold air outbreaks (i.e., periods of extreme cold lasting from several days to over a week) in NH winter in most areas, with exceptions occurring in areas with the smallest reductions of extreme cold in western North America, the North Atlantic, and southern Europe and Asia due to atmospheric circulation changes associated with the increase of GHGs.

There is evidence from modelling studies that future tropical cyclones could become more severe with greater wind speeds and more intense precipitation. Some modelling studies have projected a decrease in tropical cyclone numbers globally due to the increased stability of the tropical troposphere in a warmer climate, with fewer weaker storms and greater numbers of intense storms.

A number of modelling studies have projected a general tendency for more intense but fewer extratropical storms, with a tendency towards more extreme wind events and higher waves in association with those deepened cyclones for several regions. Models are consistent in projecting a poleward shift of storm tracks in both hemispheres by several degrees latitude.

Question 10.2: How Likely are Major or Abrupt Climate Changes, such as Loss of Ice Sheets or Changes in Global Ocean Circulation?

Based on currently available results from a hierarchy of models, we conclude that abrupt climate changes, such as the loss of the Greenland ice sheet or large-scale changes of ocean circulation systems, are not likely in the 21st century, but their occurrence becomes more likely as the perturbation of the climate system progresses.

Physical, chemical and biological analyses from Greenland ice cores, marine sediments from the North Atlantic and elsewhere, and many other paleoclimatic archives have demonstrated that local temperature, wind regimes, and the water cycle can change rapidly within just a few years. The comparison of results from records in different locations of the world shows that in the past these were major changes of hemispheric to global extent. This has led to the notion of an unstable climate in the past which underwent phases of abrupt swings. Therefore, an important question we must address is whether the continuing growth of greenhouse gas concentrations in the atmosphere constitutes a perturbation sufficiently strong to trigger abrupt changes in the climate system. If this is the case, such interference with the climate system could be considered dangerous, because it would have major global consequences.

Before the discussion of the different possibilities of such changes, we must say what we mean by "abrupt" and "major". There is no rigorous definition, but "abrupt" conveys the meaning that the changes occur much faster than the perturbation which is inducing the change. In the case of greenhouse gases, which increase on a time scale of several decades to centuries, "abrupt" would denote changes that evolve on a few decades or less. For example, if the extension of the Gulf Stream in the Atlantic Ocean, which delivers substantial amounts of heat into the northern latitudes, changed course, or if deepwater formation in the North Atlantic even ceased within a few years in response to the continuing warming, this would clearly be considered an abrupt change. By "major" we mean that the climate change involves changes that exceed the range of current natural variability, and whose spatial extent is several 1000 km, hemispheric, or global.

On more regional scales, abrupt changes are a common characteristic of natural climate variability. Here, we do not consider isolated, short lived events which are more appropriately referred to as "extreme events", but rather changes that evolve rapidly and are to stay for several years to decades. For instance, the shift in sea surface temperatures in the Eastern Pacific of the mid 1970's, or the freshening of the upper 1000 meters of the Labrador Sea since the mid 1980ies are typical examples of abrupt events with local to regional consequences.

In the public, the most widely discussed abrupt change is the collapse, or shut-down of the Gulf Stream. The Gulf Stream is a primarily horizontal current driven by the zonal wind system, which is ultimately caused by the large-scale meridional temperature contrasts at the surface of the rotating Earth. Although variable through millennia, these temperature contrasts are fundamental to the energy balance of the Earth, and therefore the Gulf Stream is a stable feature of the general circulation of the ocean. However, its northern extension which feeds deep water formation in the Greenland-Norwegian-Iceland Seas and thereby delivers heat to these areas, is shown to be strongly influenced by changes in density of the surface waters in these areas. This current constitutes the northern end of a basin-scale meridional overturning circulation (MOC) which is established along the western boundary of the Atlantic basin. A firm result of the entire hierarchy of climate models is that if the density of the surface waters in the North Atlantic decreases by warming or freshening, the MOC is reduced, and with it the delivery of heat into these areas. Strong sustained freshening induces substantial reductions, or complete shut-downs, of the MOC in all climate models. Such changes have indeed happened in the past. According to the paleoclimatic records, the MOC has shut down repeatedly during the last ice age, likely in response to massive discharges of ice from the circum-Atlantic ice sheets.

The issue now is, whether anthropogenic forcing constitutes a strong enough perturbation to the MOC that such a change can be induced. The increase of greenhouse gases in the atmosphere leads to a warming and an intensification of the hydrological cycle, with the latter making the surface waters in the Atlantic fresher. Both effects would reduce the density of the surface waters and lead to a reduction of the MOC in the 21st century. This reduction is predicted to proceed in lockstep with the warming: none of the current models simulates an abrupt reduction or shut-down. There is still a large spread in the simulated reduction of the

MOC, ranging from virtually no response to a reduction of 60% by the end of the 21st century. This is due to various strengths of the ocean, and atmosphere-ocean feedback mechanisms simulated in these models.

Uncertainty also exists about the long-term fate of the MOC. Many models show a recovery of the MOC once the warming is stabilized. But some models have thresholds of the MOC, and they are passed when the forcing is strong enough and lasts long enough. Such simulations then show a slow spin-down of the MOC which continues after the warming is stabilized. A quantification of likelihood for a long-term spin-down of the Atlantic MOC is not possible at this stage. However, even if such a scenario were to happen, Europe will still experience warming since radiative forcing overwhelms the cooling associated with the MOC reduction. In consequence, catastrophic scenarios about the beginning of an ice age triggered by a shut-down of the MOC are mere speculations, and no climate model has produced such an outcome. In fact, the processes leading to an ice age are sufficiently well understood and completely different from those discussed here, that we can confidently exclude this scenario.

Irrespective of the long term evolution of the MOC, model simulations agree that the warming and freshening will reduce deep and intermediate water formation in the Labrador Sea significantly during the next few decades. This will alter the characteristics of the intermediate water masses in the North Atlantic and eventually affect the deep ocean. The long term effects of such a change are unknown.

Other widely discussed examples of climate surprises are the loss of the Greeland ice sheet, or the abrupt disintegration of the West Antarctic Ice Sheet (WAIS). Model simulations indicate that warming in the high latitudes of the northern hemisphere accelerates the melting of the Greenland ice sheet, and that the intensified hydrological cycle is unable to compensate for this. In consequence, the Greenland ice sheet may reduce its size substantially in the coming centuries. Moreover, the results also suggest that there may exist a critical warming beyond which, if sustained, the Greenland ice sheet may disappear completely. The reduction of the Greenland ice sheet, however, is a slow process taking many hundreds of years to complete.

Recent satellite and in situ observations of ice streams behind disintegrating ice shelves highlight the rapid but transient reaction of such systems. This raises new concern about the overall stability of the West Antarctic Ice Sheet. While these streams appear buttressed by the shelves before them, it is currently unknown whether a reduction or failure of this buttressing of relatively limited areas of the ice sheet could actually trigger a wide spread discharge of many ice streams and hence a destabilization of the entire WAIS. Ice sheet models are only beginning to capture such small-scale dynamical processes which involve complicated interactions with the glacier bed and the ocean at the perimeter of the ice sheet. Therefore, we have no quantitative information available from the current generation of ice sheet models as to the likelihood of such an event.

Question 10.3: If Emissions of Greenhouse Gases are Reduced, How Quickly do Their Concentrations in the Atmosphere Decrease?

This depends crucially on the natural cycles and processes of a greenhouse gas in the climate system. Concentrations could change almost immediately in response to the emission change, or they could actually continue to increase for centuries.

The concentration of a greenhouse gas in the atmosphere depends primarily on the rate of emissions of this gas, but also on the physical and biogeochemical processes that remove it from the atmosphere. For example, carbon dioxide (CO_2) is exchanged between the atmosphere, the ocean and the land through processes such as atmosphere-ocean gas transfer and chemical (e.g., weathering) and biological (e.g., photosynthesis) processes. Methane (CH_4) is removed by chemical processes in the atmosphere, while nitrous oxide (N_2O) and some halocarbons are destroyed in the upper atmosphere by solar radiation. These processes each operate at different time scales ranging from years to millennia. A measure for this is the *lifetime* of a gas in the atmosphere, defined as the time it takes for a perturbation to reduce to 37% of its initial value. While for CH_4, N_2O, and other trace gases such as HCFC-22, a refrigerant fluid, such lifetimes can be reasonably determined (for CH_4 it is ≈12 yr, for N_2O ≈120 yr, for HCFC-22 ≈12 yr), a lifetime for CO_2 cannot be defined. Some fraction of emitted CO_2 will continue to remain in the atmosphere for many millennia owing to adjustments in the chemical composition of the surface waters of the world ocean (the ocean becomes more acid when atmospheric CO_2 increases).

The change in concentration of any trace gas will then depend in part on the time evolution of its emissions. If emissions increase with time, the atmospheric concentration will also increase with time, regardless of the atmospheric lifetime of the gas. However, if actions are taken to reduce the emissions, the fate of the trace gas concentration will depend on the relative changes not only of emissions but also of its removal processes. Here we show how the lifetimes and removal processes of different gases dictate the evolution of concentrations when emissions are reduced.

As examples, Question 10.3, Figure 1 shows test cases illustrating how the future concentration of three trace gases would respond to illustrative changes in emissions. We consider CO_2, with no specific lifetime, a trace gas with a well-defined long lifetime of the order of a century (e.g., N_2O), and a trace gas with a well-defined short lifetime of the order of decade (such as CH_4 , HCFC-22, or other halocarbons). For each gas, five idealized illustrative cases of future emissions are presented: stabilization of emissions at present-day levels, and immediate emission reduction by 10%, 30%, 50% and 100%.

[INSERT QUESTION 10.3, FIGURE 1 HERE]

The behaviour of CO_2 (Question 10.3, Figure 1a) is completely different from the trace gases with well-defined lifetimes. Stabilization of CO_2 emissions to current levels would result in a continuous increase of atmospheric CO_2 over the 21st century and beyond, whereas for a gas with a lifetime of the order of a century (Question 10.3, Figure 1b), or decade (Question 10.3, Figure 1c), stabilization of emissions to current levels would lead to a stabilization of its concentration at level higher than today within a couple of centuries, or decades, respectively. In fact, only in the case of complete reduction of emissions can the atmospheric concentration of CO_2 eventually be stabilized at a constant level. All other cases of CO_2 emission reductions show increasing concentrations because of the characteristic exchange processes associated with the cycling of carbon in the climate system.

More specifically, the rate of emission of CO_2 currently greatly exceeds its rate of removal, and the slow and incomplete removal implies that reductions in its emissions would not result in stabilization of CO_2 concentrations, but rather would reduce the rate of its growth in coming decades. A 10% reduction in CO_2 emissions would be expected to reduce the growth rate by 10%, while a 30% reduction in emissions would similarly reduce the growth rate of atmospheric CO_2 concentrations by 30%. A 50% reduction would stabilize atmospheric CO_2, but only for less than a decade, then atmospheric CO_2 would be expected to rise again as the land and ocean sinks decline owing to well-known chemical and biological adjustments. The extreme case of a complete elimination of the CO_2 emissions is estimated to lead to a slow decrease of atmospheric CO_2 of about 40ppm over the 21st century.

The situation is completely different for the trace gases with a well-defined lifetime. For the illustrative trace gas with a lifetime of the order of a century (e.g., N_2O), emission reduction of more than 50% is required to stabilized the concentrations close to present-day values (Question 10.3, Figure 1b). Constant emission leads to a stabilization of the concentration within a few centuries.

In the case of the illustrative gas with the short lifetime, the present-day loss is around 70% of the emissions. A reduction in emissions of less than 30% would still produce a transient increase in concentration in this case, but, in contrast to CO_2, would lead to stabilization of its concentration within a couple of decades (Question 10.3, Figure 1c). The decrease of the level, at which the concentration of such a gas would stabilize, is directly proportional to the emission reduction. Thus, in this illustrative example, a reduction of emission of this trace gas larger than 30% would be required to stabilize concentrations at levels significantly below those at present. A complete cut-off of the emission would lead to a return to pre-industrial concentrations within less than a century for a trace gas with a lifetime of the order of a decade.