Chapter 8: Climate Models and Their Evaluation

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Supplementary Material

The following supplementary material is available on CDRom and in on-line versions of this report.

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Executive Summary

This chapter assesses the capacity of the global climate models used elsewhere in this report for projecting future climate change. Confidence in model estimates of future climate evolution has been enhanced via a range of advances since the TAR.

Climate models are based on well established physical principles and have been demonstrated to reproduce observed features of recent climate (see Chapters 8, 9) and past climate changes (see Chapter 6). There is considerable confidence that AOGCMs provide credible quantitative estimates of future climate change, particularly at continental scales and above. Confidence in these estimates is higher for some climate variables (e.g., temperature) than for others (e.g., precipitation). In this summary we highlight areas of progress since the TAR:

- Enhanced scrutiny of models and expanded diagnostic analysis of model behavior has been increasingly facilitated by internationally coordinated efforts to collect and disseminate output from model experiments performed under common conditions. This has encouraged a more comprehensive and open evaluation of models. The expanded evaluation effort, encompassing a diversity of perspectives, makes it less likely that significant model errors are being overlooked.
- Climate models are being subjected to more comprehensive tests, including for example evaluations of forecasts on time scales from days to a year. This more diverse set of tests increases confidence in the fidelity with which models represent processes that impact climate projections.
- Substantial progress has been made in understanding the inter-model differences in equilibrium climate sensitivity. Cloud feedbacks have been confirmed as a primary source of these differences, with low clouds making the largest contribution. New observational and modelling evidence strongly supports a combined water vapour-lapse rate feedback of a strength comparable to that found in GCMs (approximately 1 W m^{-2} K^{-1}, corresponding to around a 50% amplification of global mean warming). The magnitude of cryospheric feedbacks remains uncertain, contributing to the range of model climate responses at mid-to-high latitudes.
- There have been ongoing improvements to resolution, computational methods and parametrisations, and additional processes (e.g., interactive aerosols) have been included in more of the climate models.
- Most AOGCMs no longer use flux adjustments, which were previously required to maintain a stable climate. At the same time there have been improvements in the simulation of many aspects of present climate. The uncertainty associated with the use of flux adjustments has therefore decreased, although biases and long term trends remain in AOGCM control simulations.
- Progress in the simulation of important modes of climate variability has increased our overall confidence in the models' representation of important climate processes. As a result of steady progress, some AOGCMs can now simulate important aspects of ENSO. Simulation of the MJO remains unsatisfactory.
- The ability of AOGCMs to simulate extreme events, especially hot and cold spells, has improved. The frequency and amount of precipitation falling in intense events are underestimated.
- Simulation of extratropical cyclones has improved. Some models used for projections of tropical cyclone changes can simulate successfully the observed frequency and distribution of tropical cyclones.
- Systematic biases have been found in most models' simulation of the Southern Ocean. Since the Southern Ocean is important for ocean heat uptake this results in some uncertainty in transient climate response.
- The possibility that metrics based on observations might be used to constrain model projections of climate change has been explored for the first time, through the analysis of ensembles of model simulations. Nevertheless, a proven set of model metrics that might be used to narrow the range of plausible climate projections has yet to be developed.
- To explore the potential importance of carbon cycle feedbacks in the climate system, explicit treatment of the carbon cycle has been introduced in a few climate AOGCMs and some Earth System Models of Intermediate Complexity (EMICs).
- EMICs have been evaluated in greater depth than previously. Coordinated intercomparisons have demonstrated that these models are useful in addressing questions involving long timescales or requiring a large number of ensemble simulations or sensitivity experiments.

Developments in model formulation

Improvements in atmospheric models include reformulated dynamics and transport schemes, and increased horizontal and vertical resolution. Interactive aerosol modules have been incorporated into some models, and through these, the direct and the indirect effects of aerosols are now more widely included.

Significant developments have occurred in the representation of terrestrial processes. Individual components continue to be improved via systematic evaluation against observations and against more comprehensive models. The terrestrial processes that might significantly affect large-scale climate over the next few decades are included in current climate models. Some processes important on longer time scales are not yet included.

Development of the oceanic component of AOGCMs has continued. Resolution has increased and models have generally abandoned the "rigid lid" treatment of the ocean surface. New physical parameterizations and numerics include true freshwater fluxes, improved river and estuary mixing schemes, and the use of positive definite advection schemes. Adiabatic isopycnal mixing schemes are now widely used. Some of these improvements have led to a reduction in the uncertainty associated with the use of less sophisticated parameterizations (e.g., virtual salt flux).

Progress in developing AOGCM cryospheric components is clearest for sea ice. Almost all state-of-the-art AOGCMs now include more elaborate sea-ice dynamics and some now include several sea-ice thickness categories and relatively advanced thermodynamics. AOGCM parameterizations of terrestrial snow processes vary considerably in formulation. Systematic evaluation of snow suggests that sub-grid scale heterogeneity is important for simulating observations of seasonal snow cover. Few AOGCMs include ice sheet dynamics; in all of the AOGCMs evaluated in this chapter and used in Chapter 10 for projecting climate change in the 21st Century, the land ice cover is prescribed.

There is currently no consensus on the optimal way to divide computer resources between: finer numerical grids, which allow for better simulations; greater number of ensemble members, which allow for better statistical estimates of uncertainty; and inclusion of a more complete set of processes (e.g., carbon feedbacks, atmospheric chemistry interactions).

Developments in model climate simulation

The large-scale patterns of seasonal variation in several important atmospheric fields are now better simulated by AOGCMs than they were at the time of the TAR. Notably, errors in simulating the monthly mean, global distribution of precipitation, sea level pressure, and surface air temperature have all decreased. In some models, simulation of marine low-level clouds, which are important for correctly simulating sea surface temperature and cloud feedback in a changing climate, has also improved. Nevertheless, important deficiencies remain in the simulation of clouds and tropical precipitation (with their important regional and global impacts).

Some common model biases in the Southern Ocean have been identified, resulting in some uncertainty in oceanic heat uptake and transient climate response. Simulations of the thermocline, which was too thick, and the Atlantic overturning and heat transport, which were both too weak, have been substantially improved in many models.

Despite notable progress in improving sea ice formulations, AOGCMs have typically achieved only modest progress in simulations of observed sea-ice since the TAR. The relatively slow progress can partially be explained by the fact that improving sea ice simulation requires improvements in both the atmosphere and ocean components in addition to the sea ice component itself.

Since the TAR, developments in AOGCM formulation have improved the representation of large-scale variability over a wide range of time-scales. The models capture the dominant extratropical patterns of variability including the Northern and Southern Annular Modes, the Pacific Decadal Oscillation, the Pacific-North American and Cold Ocean-Warm Land Patterns. AOGCMs simulate Atlantic multidecadal variability, although the relative roles of high and low latitude processes appear to differ between models. In the tropics, there has been an overall improvement in the AOGCM simulation of the spatial pattern and frequency of the

El Niño – Southern Oscillation, but problems remain in simulating its seasonal phase locking and the asymmetry between El Niño and La Niña episodes. Variability with some characteristics of the Madden-Julian Oscillation is simulated in most AOGCMs, but the events are typically too infrequent and too weak.

AOGCMs are able to simulate extreme warm temperatures, cold air outbreaks and frost days reasonably well. Models used in this report for projection of tropical cyclone changes are able to simulate present day frequency and distribution of cyclones, but intensity is less well simulated. Simulation of extreme precipitation is dependent on resolution, parametrization and the thresholds chosen. In general models tend to produce too many days with weak precipitation (<10 mm day^{-1}) and too little precipitation overall in intense events (>10 mm day^{-1}).

Earth system models of intermediate complexity (EMICs) have been developed to investigate issues in past and future climate change that cannot be addressed by comprehensive AOGCMs because of their large computational cost. Owing to the reduced resolution of EMICs and their simplified representation of some physical processes, these models only allow inferences about very large scales. Since the TAR, EMICs have been evaluated via several coordinated model intercomparisons which have revealed that, at large scales, EMIC results can compare well with observational data and AOGCM results. This lends support to the view that EMICS can be used to gain understanding of processes and interactions within the climate system that evolve on time-scales beyond those generally accessible to current AOGCMs. The uncertainties in long-term climate change projections can also be explored more comprehensively by using large ensembles of EMIC runs.

Developments in analysis methods

Since the TAR, an unprecedented effort has been initiated to make available new model results for scrutiny by scientists outside the modelling centers. Eighteen modeling groups performed a set of coordinated, standard experiments, and the resulting model output, analyzed by hundreds of researchers worldwide, forms the basis for much of the current IPCC assessment of model results. The benefits of coordinated model intercomparison include increased communication among modelling groups, more rapid identification and correction of errors, the creation of standardized benchmark calculations, and a more complete and systematic record of modelling progress.

A few climate models have been tested for (and shown) capability in initial value predictions, on timescales from weather forecasting (a few days) to seasonal forecasting (annual). The capability demonstrated by models under these conditions increases confidence that they simulate some of the key processes and teleconnections in the climate system.

Developments in evaluation of climate feedbacks

Water vapour feedback is the most important feedback enhancing climate sensitivity. Although the strength of this feedback varies somewhat among models, its overall impact on the spread of model climate sensitivities is reduced by lapse rate feedback, which tends to be anticorrelated. Several new studies indicate that modelled lower and upper tropospheric humidity respond to seasonal and interannual variability, volcanically induced cooling and climate trends, in a way that is consistent with observations. Recent observational and modelling evidence thus provides strong additional support for the combined water vapour-lapse rate feedback being around the strength found in AOGCMs.

Recent studies reaffirm that the spread of climate sensitivity estimates among models arises primarily from inter-model differences in cloud feedbacks. The shortwave impact of changes in boundary-layer clouds, and to a lesser extent mid-level clouds, constitutes the largest contributor to inter-model differences in global cloud feedbacks. The relatively poor simulation of these clouds in the present climate is a reason for some concern. The response to global warming of deep convective clouds is also a substantial source of uncertainty in projections since current models predict different responses of these clouds. Observationally-based evaluation of cloud feedbacks indicates that climate models exhibit different strengths and weaknesses, and it is not yet possible to determine which estimates of the climate change cloud feedbacks are the most reliable.

Despite advances since the TAR, substantial uncertainty remains in the magnitude of cryospheric feedbacks within AOGCMs. This contributes to a spread of modelled climate response, particularly in high latitudes. On the global scale, the surface albedo feedback is positive in all the models, and varies between models much less than cloud feedbacks. Understanding and evaluating sea-ice feedbacks is complicated by the strong coupling to polar cloud processes and ocean heat and freshwater transport. Scarcity of observations in polar regions also hampers evaluation. New techniques that evaluate surface albedo feedbacks have recently been developed. Model performance in reproducing the observed seasonal cycle of land snow cover may provide an indirect evaluation of the simulated snow-albedo feedback under climate change.

Systematic model comparisons have helped establish the key processes responsible for differences among models in the response of the ocean to climate change. The importance of feedbacks from surface flux changes on the meridional overturning circulation has been established in many models. At present, these feedbacks are not tightly constrained by available observations.

The analysis of processes contributing to climate feedbacks in models, and recent studies based on large ensembles of models, suggest that in the future it may be possible to use observations to narrow the current spread in model projections of climate change.

8.1 Introduction and Overview

The goal of this chapter is to evaluate the capabilities and limitations of the global climate models used elsewhere in this assessment. A number of model evaluation activities are described in various chapters of this report. This section provides a context for those studies and a guide to direct the reader to the appropriate chapters.

8.1.1 What is Meant by Evaluation?

A specific prediction based on a model can often be demonstrated to be right or wrong, but the model itself should always be viewed critically. This is true for both weather prediction and climate prediction. Weather forecasts are produced on a regular basis, and can be quickly tested against what actually happened. Over time, statistics can be accumulated that give information on the performance of a particular model or forecast system. In climate change simulations, on the other hand, we use models to make projections of possible future changes, for which timescales are many decades and for which there are no precise past analogues. We can gain confidence in a model through simulations of the historical record, or of paleoclimate, but such opportunities are much more limited than those available through weather prediction. These and other approaches are discussed below.

8.1.2 Methods of Evaluation

A climate model is a very complex system, with many components. The model must of course be tested at the system level, i.e., by running the full model and comparing the results with observations. Such tests can reveal problems, but their source is often hidden by the model's complexity. For this reason, it is also important to test the model at the component level, i.e., by isolating particular components and testing them independent of the complete model.

Component-level evaluation of climate models is common. Numerical methods are tested in standardized tests, organized through activities such as the quasi-biennial Workshops on Partial Differential Equations on the Sphere. Physical parameterizations used in climate models are being tested through numerous case studies (some based on observations and some idealized), organized through programs such as ARM, EUROCS, and GCSS. These various activities have been ongoing for a decade or more. A large body of results has been published (e.g., Randall et al., 2003).

System-level evaluation is focused on the outputs of the full model, i.e., model simulations of particular observed climate variables, and particular methods are discussed in more detail below.

8.1.2.1 Model Intercomparisons and Ensembles

The global model intercomparison activities that began in the late 1980s (e.g., Cess et al., 1989), and continued with AMIP (the Atmosphere Model Intercomparison Project), have now proliferated to include several dozen “MIPs”, covering virtually all climate model components and various coupled model configurations. A summary is available at http://www.ifm.uni-kiel.de/other/clivar/science/mips.htm. By far the most ambitious organized effort to collect and analyze AOGCM output from standardized experiments was undertaken in the last few years (see http://www-pcmdi.llnl.gov/ipcc/about_ipcc.php). It differed from previous model intercomparisons in that a more complete set of experiments was performed, including unforced control simulations, simulations attempting to reproduce observed climate change over the instrumental period, and simulations of future climate change. It also differed in that, for each experiment, multiple simulations were performed by some individual models to make it easier to separate climate change signals from internal variability within the climate system. Perhaps the most important change from earlier efforts was the collection of a more comprehensive set of model output, hosted centrally at PCMDI. This archive, referred to here as ‘The Multi-Model Dataset (MMD) at PCMDI’, has allowed hundreds of researchers from outside the modeling groups to scruitinize the models from a variety of perspectives.

The enhancement in diagnostic analysis of climate model results represents an important step forward since the TAR. Overall, the vigorous, ongoing intercomparison activities have increased communication among modelling groups, allowed rapid identification and correction of modeling errors, and encouraged the creation of standardized benchmark calculations, as well as a more complete and systematic record of modelling progress.

Ensembles of models represent a new resource for studying the range of plausible climate responses to a given forcing. Such ensembles can be generated either by collecting results from a range of models from different modelling centres ('multi-model ensembles' as described above), or by generating multiple model versions within a particular model structure, by varying internal model parameters within plausible ranges ('perturbed physics ensembles'). The approaches are discussed in mode detail in Section 10.5.

8.1.2.2 Metrics of Model Reliability

What does the accuracy of a climate model's simulation of past or contemporary climate tell us about the accuracy of its projections of climate change? This question is just beginning to be addressed, exploiting the newly available ensembles of models. A number of different observationally based metrics have been used to weight the reliability of contributing models when making probabilistic projections (see Section 10.5.4).

For any given metric, it is important to assess how good a test it is of model results for making projections of future climate change. This cannot be tested directly, since there are no observed periods with forcing changes exactly analogous to those expected over the 21st Century. However, relationships between observable metrics and the predicted quantity of interest (e.g. climate sensitivity) can be explored across model ensembles. In Shukla et al. (2006), a measure of the fidelity of the simulated surface temperature in the 20th Century was correlated with simulated 21st Century temperature change in a multi-model ensemble. It was found that the models with the smallest 20th Century error produced relatively large surface temperature increases in the 21st Century. Knutti et al. (2006), using a different, perturbed physics ensemble, showed that models with a strong seasonal cycle in surface temperature tended to have larger climate sensitivity. More complex metrics have also been developed based on multiple observables in present day climate, and have been shown to have the potential to narrow the uncertainty in climate sensitivity across a given model ensemble (Murphy et al., 2004; Piani et al., 2005). The above studies show promise that quantitative metrics for the likelihood of model projections may be developed, but because the development of robust metrics is still at an early stage, the model evaluations presented in this chapter are based primarily on experience and physical reasoning, as has been the norm in the past.

An important area of progress since the TAR has been in establishing and quantifying the feedback processes that determine climate change response. Knowledge of these processes underpins both the traditional and the metric-based approaches to model evaluation. For example, in Hall and Qu (2006), a metric was developed for the feedback between temperature and albedo in snow covered regions, based on the simulation of the seasonal cycle. It was found that models with a strong feedback, based on the seasonal cycle, also had a strong feedback under increased greenhouse gas forcing. Comparison with observed estimates of the seasonal cycle suggested that most models in the MMD underestimate the strength of this feedback. Section 8.6 discusses the various feedbacks which operate in the atmosphere-land surface-sea ice system to determine climate sensitivity, and Section 8.3.2 discusses some processes that are important for ocean heat uptake (and hence transient climate response).

8.1.2.3 Testing Models Against Past and Present Climate

Testing models' ability to simulate 'present climate' (including variability and extremes) is an important part of model evaluation (see Sections 8.3 to 8.5, and Chapter 11 for specific regional evaluations). In doing this, certain practical choices are needed, e.g. between a long timeseries or mean from a 'control' run with fixed radiative forcing (often preindustrial rather than present day), or a shorter, transient timeseries from a '20th-century' simulation including historical variations in forcing. Such decisions are made by individual researchers, dependent on the particular problem being studied. Differences between model and observations should be considered insignificant if they are within

1. unpredictable internal variability (e.g., the observational period contained an unusual number of El Niño events)
2. expected differences in forcing (e.g., observations for the 1990s compared with a 'preindustrial' model control run)

3. uncertainties in the observed fields

and while space does not allow a discussion of the above issues in detail for each climate variable, they are taken into account in the overall evaluation. Model simulation of present day climate on a global to subcontinental scale is discussed in Chapter 8, while more regional detail can be found in Chapter 11.

Models have been extensively used to simulate observed climate change during the 20th Century. Since forcing changes are not perfectly known over that period (see Chapter 2), such tests do not fully constrain future response to forcing changes. Knutti et al. (2002) show that in a perturbed-physics EMIC ensemble, models with a range of climate sensitivities are consistent with the observed surface air temperature and ocean heat content records, if aerosol forcing is allowed to vary within its range of uncertainty. Despite this fundamental limitation, testing of 20th Century simulations against historical observations does place some constraints on future climate response (e.g., Knutti et al. 2002). These topics are discussed in detail in Chapter 9.

8.1.2.4 Other Methods of Evaluation

Simulations of climate states from the more distant past allow models to be evaluated in regimes that are significantly different from the present. Such tests complement the 'present climate' and 'instrumental period climate' evaluations, since 20th Century climate variations have been small compared with the anticipated future changes under SRES forcing scenarios. The limitations of palaeoclimate tests are that uncertainties in both forcing and actual climate variables (usually derived from proxies) tend to be greater than in the instrumental period, and that the number of climate variables for which there are good palaeo-proxies is limited. Further, climate states may have been so different (e.g., ice sheets at last glacial maximum) that processes determining quantities such as climate sensitivity were different from those likely to operate in the 21st Century. Finally, the timescales of change were so long that there are difficulties in experimental design, at least for GCMs. These issues are discussed in depth in Chapter 6.

Climate models can be tested through forecasts based on initial conditions. Climate models are closely related to the models that are used routinely for numerical weather prediction (NWP), and increasingly for extended range forecasting on seasonal to interannual timescales. Typically, however, models used for NWP are run at higher resolution than is possible for climate. Evaluation of such forecasts tests the models' representation of some key processes in the atmosphere and ocean, although the links between these processes and long-term climate response have not always been established. It must be remembered that the quality of an initial value prediction is dependent on several factors beyond the numerical model itself (e.g., data assimilation techniques, ensemble generation method), and these factors may be less relevant to projecting the long term, forced response of the climate system to changes in radiative forcing. There is a large literature on this topic, but to maintain focus on the goal of this chapter we confine ourselves to the relatively few studies that have been conducted using models that are very closely related to the climate models used for projections (see Section 8.4.11).

8.1.3 How Are Models Constructed?

The fundamental basis on which climate models are constructed has not changed since the TAR, although there have been many specific developments (see Section 8.2). Climate models are derived from fundamental physical laws (such as Newton's laws of motion), which are then subjected to physical approximations appropriate for the large-scale climate system, and then further approximated through mathematical discretization. Computational constraints restrict the resolution that is possible in the discretised equations, and some representation of the large-scale impacts of unresolved processes is required (the parametrisation problem).

8.1.3.1 Parameter Choices and 'Tuning'

Parameterizations are typically based in part on simplified physical models of the unresolved processes (e.g., entraining plume models in some convection schemes). The parameterizations also involve numerical parameters that must be specified as input. Some of these parameters can be measured, at least in principle, while others cannot. It is therefore common to adjust parameter values (possibly chosen from some prior distribution) in order to optimise model simulation of particular variables or to improve global heat balance. This process is often known as tuning. It is justifiable to the extent that two conditions are met:

1. Observationally-based constraints on parameter ranges are not exceeded. Note that in some cases this may not provide a tight constraint on parameter values (e.g., Heymsfield and Donner, 1990).

2. The number of degrees of freedom in the tunable parameters is less than the number of degrees of freedom in the observational constraints used in model evaluation. This is believed to be true for most GCMs – for example climate models are not explicitly tuned to give a good representation of NAO variability – but no studies are available that address the question formally. If the model has been tuned to give a good representation of a particular observed quantity, then agreement with that observation cannot be used to build confidence in that model. However, a model that has been tuned to give a good representation of certain key observations may have a greater likelihood of giving a good prediction than a similar model (perhaps another member of a 'perturbed physics' ensemble) which is less closely tuned (as discussed in 8.1.2.2 above and in Chapter 10).

Given sufficient computer time the tuning procedure can in principle be automated using various data assimilation procedures. To date, however, this has only been feasible for EMICs (Hargreaves et al., 2004) and low-resolution GCMs (Annan et al., 2005b; Jones et al., 2005; Severijns and Hazeleger, 2005). Ensemble methods (Murphy et al., 2004; Annan et al., 2005a; Stainforth et al., 2005) do not always produce a unique 'best' parameter setting for a given error measure.

8.1.3.2 Model Spectra or Hierarchies

The value of using a range of models (a 'spectrum' or 'hierarchy') of differing complexity is discussed in the TAR (Section 8.3), and here in Section 8.8. Computationally cheaper models such as EMICs allow a more thorough exploration of parameter space, and are simpler to analyse to gain understanding of particular model responses. Models of reduced complexity have been used more extensively in this report than in the TAR, and their evaluation is discussed in Section 8.8.We note that regional climate models can also be viewed as forming part of a climate-modeling hierarchy.

8.2 Advances in Modelling

Many modeling advances have occurred since the TAR. Space does not permit a comprehensive discussion of all major changes made over the past several years to the twenty-three AOGCMs used widely in this report (see Table 8.1). Model improvements can, however, be grouped into three categories. First, the dynamical cores (advection etc.) have been improved, and the horizontal and vertical resolution of many models have been increased. Second, more processes have been incorporated into the models, in particular in the modelling of aerosols, and of land-surface and sea-ice processes. Third, the parameterizations of physical processes have been improved. For example, as discussed further in Section 8.2.7, most of the models no longer use flux adjustments (Manabe and Stouffer, 1988; Sausen et al., 1988) to reduce climate drift. These various improvements, developed across the broader modelling community, are well represented in the climate models used in this report.

Despite the many improvements, numerous issues remain. Many of the important processes that determine a model's response to changes in radiative forcing are not resolved by the model's grid. Instead subgrid scale parameterizations are used to parameterize the unresolved processes, such as cloud formation and the mixing due to oceanic eddies. It continues to be the case that multi-model ensemble simulations generally provide more robust information than runs of any single model. Refer to Table.8.1 for a summary of the formulations of each of the AOGCMs used in this report.

There is currently no consensus on the optimal way to divide computer resources between: finer numerical grids, which allow for better simulations; greater number of ensemble members, which allow for better statistical estimates of uncertainty; and inclusion of a more complete set of processes (e.g., carbon feedbacks, atmospheric chemistry interactions).

[INSERT TABLE 8.1 HERE]

8.2.1 Atmospheric Processes

8.2.1.1 Numerics

In the TAR, more than half of the participating atmospheric models used spectral advection. Since the TAR, semi-Lagrangian advection schemes have been adopted in several atmospheric models. These schemes allow long time steps and maintain positive values of advected tracers such as water vapor, but they are diffusive, and some versions do not formally conserve mass. In this report, various models use spectral, semi-Lagrangian, and Eulerian finite-volume and finite-difference advection schemes ,although there is still no consensus on which type of scheme is best.

8.2.1.2 Horizontal and Vertical Resolution

The horizontal and vertical resolutions of AOGCMs have increased relative to the TAR. For example, HadGEM1 has 8 times as many grid cells as HadCM3 (the number of cells has doubled in all three dimensions). At NCAR, a T85 version of the CSM is now routinely used, while a T42 version was standard at the time of the TAR. CCSR-NIES-FRCGC has developed a high-resolution climate model (MIROC-hi, which consists of a T106L56 AGCM and a 1/4° by 1/6° L48 OGCM), and MRI/JMA has developed a TL959 L60 spectral AGCM (Oouchi et al., 2006), which is being used in time-slice mode. The projections made with these models are presented in Chapter 10.

Due to the increased horizontal and vertical resolution, both regional and global-scale climate features are better simulated. For example, a far-reaching effect of the Hawaiian Islands in the Pacific Ocean (Xie et al., 2001) has been well simulated (Sakamoto et al., 2004) and the frequency-distribution of precipitation associated with the Baiu-front is improved (Kimoto et al., 2005).
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8.2.1.3 Parameterisations

The climate system includes a variety of physical processes, such as cloud processes, radiative processes and boundary-layer processes, which interact with each other on many temporal and spatial scales. Due to the limited resolutions of the models, many of these processes are not resolved adequately by the model grid and must therefore be parameterized.. The differences between parametrizations are an important reason why climate model results differ. For example, a new boundary layer parameterization (Lock et al., 2000; Lock, 2001) had a strong positive impact on the simulations of marine stratocumulus cloud produced by the GFDL climate models and the Hadley Centre, but the same parameterization had less positive impact when implemented in an earlier version of the Hadley Centre model (Martin et al., 2006). Clearly, parametrizations must be understood in the context of their host models.

Cloud processes affect the climate system by regulating the flow of radiation at the top of the atmosphere, by producing precipitation, by accomplishing rapid and sometimes deep redistributions of atmospheric mass, and through additional mechanisms too numerous to list here (Arakawa and Schubert, 1974; Arakawa, 2004). Cloud parameterizations are physically based on theories that aim to describe the statistics of the cloud field, e.g., the fractional cloudiness or the area-averaged precipitation rate, without describing the individual cloud elements. In an increasing number of climate models, microphysical parametrizations that represent such processes as cloud particle and a raindrop formation are used to predict the distributions of liquid and ice clouds. These parametrizations improve the simulation of the present climate, and affect climate sensitivity (Iacobellis et al., 2003). Realistic parameterizations of cloud processes are a prerequisite for reliable current and future climate simulation (see Section 8.6).

Data from field experiments such as GATE (1974), MONEX (1979), ARM (1993), and TOGA-COARE (1993) have been used to test and improve parameterizations of clouds and convection (e.g., Emanuel and Zivkovic-Rothmann, 1999; Sud and Walker, 1999; Bony and Emanuel, 2001). Systematic research such as that conducted by the GEWEX (Global Energy and Water Experiment) Cloud Systems Study (GCSS; Randall et al., 2003) has been organized to test parametrizations by comparing results with both observation and the results of a cloud-resolving model. These efforts have influenced the development of many of the recent models. For example, the boundary-layer cloud parameterization of Lock et al. (2000) and Lock

(2001), was tested via GCSS. Parameterizations of radiative processes have been improved and tested by comparing results of radiation parameterizations used in AOGCMs with those of much more detailed "line-by-line" radiation codes (Collins et al., 2006). Since the TAR, improvements have been made in several models to the physical coupling between cloud and convection parameterizations, e.g., in the MPI AOGCM using Tompkins (2002), in the IPSL-CM4 OAGCM using Bony and Emanuel (2001) and in the GFDL model using Tiedtke (1993). These are examples of component-level testing.

In parallel with improvement in parameterizations, a non-hydrostatic model has been used for downscaling. MRI/JMA has run a model with a 5 km grid on a domain of 4000 km by 3000 km by 22 km centered over Japan, using the time-slice method for AR4 (Yoshizaki et al., 2005).

Aerosols play an important role in the climate system. Interactive aerosol parameterizations are now used in some models (HADGEM1, MIROC-hi, MIROC-med). Both the 'direct' and 'indirect' aerosol effects (Chapter 2) have been incorporated in some cases (e.g., IPSL-CM4). In addition to sulphates, other types of aerosols such as black and organic carbon, sea-salt, and mineral dust are being introduced as prognostic variables (Takemura et al., 2005; see Chapter 2). Further details are given in Section 8.2.5.

8.2.2 Ocean Processes

8.2.2.1 Numerics

Recently, isopycnic or hybrid vertical coordinates have been adopted in some ocean models (GISS-EH and BCCR-BCM2.0). Tests show that such models can produce solutions for complex regional flows that are as realistic as those obtained with the more common depth-coordinate (e.g., Drange et al., 2005). Issues remain over the proper treatment of thermobaricity (non-linear relationship of temperature, salinity and pressure to density), which means that in some isopycnic coordinate models the relative densities of, say, Mediterranean and Antarctic Bottom Water masses are distorted. The merits of these vertical coordinate systems are still being established.

An explicit representation of the sea-surface height is being used in many models, and real freshwater flux is used to force those models instead of a "virtual" salt flux. The virtual salt flux method induces a systematic error in sea surface salinity prediction and causes a serious problem at large river basin mouths (Hasumi, 2002a,b; Griffies, 2004).

Generalized curvilinear horizontal coordinates with bipolar or tripolar grids (Murray, 1996) have become widely used in the oceanic component of AOGCMs. These are strategies used to deal with the North Pole coordinate singularity, as alternatives to the previously common polar filter or spherical coordinate rotation. The newer grids have the advantage that the singular points can be shifted onto land while keeping grid points aligned on the equator. The older methods of representing the ocean surface, surface water flux and North Pole are still in use in several AOGCMs.

8.2.2.2 Horizontal and Vertical Resolution

There has been a general increase in resolution since the TAR, with a horizontal resolution of order 1–2 degrees now commonly used in the ocean component of most climate models. To better resolve the equatorial waveguide, several models use enhanced meridional resolution in the tropics. Resolution high enough to allow oceanic eddies, eddy-permitting, has not been used in a full suite of climate scenario integrations due to computational cost, but since the TAR it has been used in some idealised and scenario-based climate experiments as discussed below. A limited set of integrations using the eddy-permitting MIROC3.2 (hires) model is used here and in Chapter 10. Some modelling centres have also increased vertical resolution since the TAR.

A few coupled climate models with eddy-permitting ocean resolution (1/6 to 1/3 degree) have been developed (Roberts et al., 2004; Suzuki et al., 2005), and large-scale climatic features induced by local air-sea coupling have been successfully simulated (e.g., Sakamoto et al., 2004).

Roberts et al. (2004) found that increasing the ocean resolution of the HadCM3 model from about 1° to 0.33° by 0.33° by 40 levels (while leaving the atmospheric component unchanged) resulted in many improvements in the simulation of features of the ocean circulation. However the impact on the atmospheric simulation was relatively small and localized. The climate change response was similar to the standard resolution model, with a slightly faster rate of warming in the Northern Europe-Atlantic region due to differences in the Atlantic MOC response. The adjustment timescale of the Atlantic basin fresh water budget decreased from O(400 years) to O(150 years) with the higher resolution ocean, suggesting possible differences in transient MOC response on those timescales, but the mechanisms and the relative roles of horizontal and vertical resolution are not clear.

The Atlantic MOC is influenced by freshwater as well as thermal forcing. Besides atmospheric freshwater forcing, freshwater transport by the ocean itself is also important. For the Atlantic MOC, the fresh Pacific water coming through the Bering Strait could be poorly simulated on its transit to the Canadian Archipelago and the Labrador Sea (Komuro and Hasumi, 2005). These aspects are improved since the TAR in many of the models evaluated here.

Changes around continental margins are very important for regional climate change. Over these areas, climate is influenced by the atmosphere and open ocean circulation. High-resolution climate models contribute to the improvement of simulation of regional climate. For example, the location of the Kuroshio separation from the Japan islands is well simulated in the MIROC3.2 (hires) model (see Figure 8.1), which makes it possible to study a change of the Kuroshio axis in future climate (Sakamoto et al., 2005).

[INSERT FIGURE 8.1 HERE]

Guilyardi et al. (2004) suggest that ocean resolution may play only a secondary role in setting the time scale of model ENSO variability, with the dominant timescales being set by the atmospheric model provided the basic speeds of the equatorial ocean wave modes are adequately represented.

8.2.2.3 Parametrisations

In the tracer equations, isopycnal diffusion (Redi, 1982) with isopycnal layer thickness diffusion (Gent et al., 1995), including its modification by Visbeck et al. (1997), has become a widespread choice instead of a simple horizontal diffusion. This has led to improvements in the thermocline structure and meridional overturning (Böning et al., 1995; see Section 8.3.2). For vertical mixing of tracers, a wide variety of parameterizations is currently used, such as turbulence closures (e.g., Mellor and Yamada, 1982), non-local diffusivity profiles (Large et al., 1994), and bulk mixed layer models (e.g., Kraus and Turner, 1967). Representation of the surface mixed layer has been much improved due to developments in these parameterizations (see 8.3.2). Observations have shown that deep ocean vertical mixing is enhanced over rough bottom and steep slopes, and where stratification is weak (Kraus, 1990; Polzin et al., 1997; Moum et al., 2002). While there have been modelling studies indicating the significance of such inhomogeneous mixing for the MOC (e.g., Marotzke, 1997; Hasumi and Suginohara, 1999; Otterå et al., 2004; Oliver et al., 2005, Saenko and Merryfield 2005), comprehensive parameterizations for the effects and their application in coupled climate models are still to be seen.

Many of the dense waters formed by oceanic convection, which are integral to the global MOC, must flow over ocean ridges or down continental slopes. The entrainment of ambient water around these topographic features is an important process determining the final properties and quantity of the deep waters. Parameterizations for such bottom boundary layer (BBL) processes have come into use in some AOGCMs (e.g., Nakano and Suginohara, 2002; Winton et al., 1998). However the impact of the BBL representation on the coupled system is not fully understood (Tang and Roberts, 2005). Thorpe et al. (2004) study the impact of the very simple scheme used in the HadCM3 model to control mixing of overflow waters from the Nordic Seas into the North Atlantic. Although the scheme does result in a change of the subpolar water mass properties, it appears to have little impact on the simulation of the large-scale THC strength or its response to global warming.

8.2.3 Terrestrial Processes

Few multi-model analyses have been conducted of terrestrial processes included in the models in Table 8.1. However, significant advances since the TAR have been reported based on climate models that are similar to these models. Analysis of these models provides insight on how well terrestrial processes are likely included in the AR4 models.

8.2.3.1 Surface Processes

The addition of the terrestrial biosphere models that simulate changes in terrestrial carbon sources and sinks into fully-coupled climate models is at the cutting edge of climate science. The major advance in this area since the TAR is the inclusion of carbon cycle dynamics including vegetation and soil carbon cycling, although these are not yet incorporated routinely into the AOGCMs used for climate projection (see Chapter 10). The inclusion of the terrestrial carbon cycle introduces a new and potentially important feedback into the climate system on time scales of decades to centuries (see Chapters 7 and 10). These feedbacks include the responses of the terrestrial biosphere to increasing CO_2, climate change and changes in climate variability (see Chapter 7). However, many issues remain to be resolved. The magnitude of the sink remains uncertain (Cox et al., 2000; Friedlingstein et al., 2001; Dufresne et al., 2002) because it depends on climate sensitivity as well as on the response of vegetation and soil carbon to increasing CO_2 (Friedlingstein et al., 2003). The rate at which CO_2-fertilization saturates in terrestrial systems dominates the present uncertainty in the role of biospheric feedbacks. A series of studies have been conducted to explore the present modelling capacity of the response of the terrestrial biosphere rather than the response of just one or two of its components (Friedlingstein et al., 2006). This work has built on systematic efforts to evaluate the capacity of terrestrial biosphere models to simulate the terrestrial carbon cycle (Cramer et al., 2001) via intercomparison exercises. For example, Friedlingstein et al. (2006) find that in all models examined the sink is reduced in the future as the climate warms.

Other individual components of land surface processes have been improved since the TAR, such as root parameterization (Arora and Boer, 2003; Kleidon, 2004), and higher resolution river routing (Ducharne et al., 2003). Cold land processes have received considerable attention with multi-layer snowpack models now being more common (e.g. Oleson et al., 2004) as is the inclusion of soil freezing and thawing (e.g., Boone et al., 2000; Warrach et al., 2001). Sub-grid scale snow parameterizations (Liston, 2004), snow-vegetation interations (Essery et al., 2003) and the wind-redistribution of snow (Essery and Pomeroy, 2004) are more commonly considered. High-latitude organic soils have are included in some models (Wang et al., 2002). A recent advance is the coupling of ground water models into land surface schemes (Liang et al. 2003; Maxwell and Miller, 2005; Yeh and Eltahir, 2005). These have only been evaluated locally but may be adaptable to global-scales. There is also evidence emerging that regional-scale projection of warming is sensitive to the simulation of processes that operate at finer scales than current climate models resolve (Pan et al., 2004). In general, the improvements in land surface models since the TAR are based on detailed comparisons against observational data. For example, Boone et al. (2004) used the Rhone Basin to investigate how land surface models' simulate the water balance for several annual cycles compared to data from a dense observation network. They found that most land surface schemes simulate very similar total runoff and evapotranspiration but the partitioning between the various components of both runoff and evaporation varies greatly resulting in different soil water equilibrium states and simulated discharge. More sophisticated snow parameterizations led to superior simulations of basin-scale runoff.

An analysis of AMIP-2 results explored the land surface contribution to climate simulation. Henderson-Sellers et al. (2003) found a clear chronological sequence of land surface schemes (early models that excluded an explicit canopy, more recent biophysically-based models and very recent biophysically based models). Statistically significant differences in annually-averaged evaporation were identified that could be associated with the parameterization of canopy processes. Further improvements in land surface models depends on enhanced surface observations, for example, the use of stable isotopes (e.g., Henderson-Sellers et al., 2004) which allow several components of evaporation to be evaluated separately. Pitman et al. (2004) explored the impact of the level of complexity used to parameterize the surface energy balance on differences found among the AMIP-2 results. They found that quite large variations in surface energy balance complexity did not lead to systematic differences in the simulated mean, minimum or maximum temperature variance at the global scale, or in the zonal averages, indicating that these variables are not limited by uncertainties in how to parameterize the surface energy balance. This adds confidence to the use of the models in Table 8.1, as most include surface energy balance modules of more complexity that the minimum identified by Pitman et al. (2004).

While little work has been performed to assess the capability of the land surface models used in coupled climate models, the upgrading of the land surface models is gradually taking place and the inclusion of carbon into these models is a major conceptual advance. In the simulation of the present day climate, the limitations of the standard bucket hydrology model are increasingly clear (Milly and Shmakin, 2002; Henderson-Sellers et al., 2004; Pitman et al., 2004) including evidence that it overestimates the likelihood of drought (Seneviratne et al., 2002). Relatively small improvements to the land surface model, for example the inclusion of spatially variable water holding capacity and a simple canopy conductance, lead to significant improvements (Milly and Shmakin, 2002). Since most models in Table 8.1 represent the continental-scale land surface more realistically than the standard bucket hydrology scheme, and include spatially variable water holding capacity, canopy conductance etc (Table 8.1) most of these models likely capture the key contribution made by the land surface to current large-scale climate simulation. However, it is not clear how well current climate models can capture the impact of future warming on the terrestrial carbon balance. A systematic evaluation of AOGCMs with the carbon cycle represented would help increase our confidence in the contribution of the terrestrial surface resulting from future warming.

8.2.3.2 Soil Moisture Feedbacks in Climate Models

A key role of the land surface is to store soil moisture and control its evaporation. An important process, the soil moisture-precipitation feedback, has been explored extensively since the TAR, building on regionally-specific studies that demonstrated links between soil moisture and rainfall. Recent studies (e.g., Gutowski et al., 2004; Pan et al., 2004) suggest that summer precipitation strongly depends on surface processes, notably in the simulation of regional extremes. Douville (2001) showed that soil moisture anomalies affect the African monsoon while Schär et al. (2004) suggest that an active soil moisture-precipitation feedback was linked to the anomalously hot European summer in 2003.

The soil moisture-precipitation feedback in climate models had not been systematically assessed at the time of the TAR. It is associated with the strength of coupling between the land and atmosphere which is not directly measurable at the large scale in nature and has only recently been quantified in models (Dirmeyer, 2001). Koster et al. (2004) provides an assessment of where the soil moisture-precipitation feedback is regionally important during the northern hemisphere summer, by quantifying the coupling strength in a dozen atmospheric GCMs. Some similarity was seen amongst the model responses, enough to produce a multi-model average estimate of where the global precipitation pattern during the northern hemisphere summer was most strongly affected by soil moisture variations. These “hot spots” of strong coupling are found in transition regions between humid and dry areas. The models, however, also show strong disagreement in the strength of land-atmosphere coupling. A few studies have explored the differences in coupling strength. Seneviratne et al. (2002) highlight the importance of differing water-holding capacities among the models while Lawrence and Slingo (2005) explore the role of soil moisture variability and suggest that frequent soil moisture saturation and low soil moisture variability could partially explain the weak coupling strength in the HadAM3 model (note that “weak” does not imply “wrong” since the real strength of the coupling is unknown.

Overall the uncertainty in surface-atmosphere coupling has implications for the reliability of the simulated soil moisture-atmosphere feedback. It tempers our interpretation of the response of the hydrologic cycle to simulated climate change in “hot spot” regions. Note that no assessment has been attempted for seasons other than northern hemisphere summer.

Since the TAR there have been few assessments of the capacity of climate models to simulate observed soil moisture. Despite the tremendous effort to collect and homogenize soil moisture measurements at global scales (Robock et al., 2000) discrepancies between large scale estimates of observed soil moisture remain. The challenge of modelling soil moisture, that naturally varies on small scales, linked to landscape characteristics, soil processes, ground water recharge, vegetation type etc, within climate models in a way that facilitates comparison with observed data is considerable. It is not clear how to compare climate model simulated soil moisture with point-based or remotely sensed soil moisture. This makes assessing how well climate models simulate soil moisture, or the change in soil moisture, difficult.

8.2.4 Cryospheric Processes

8.2.4.1 Terrestrial Cryosphere

Ice sheet models are used in calculations of long-term warming and sea level scenarios, though they have not generally been incorporated in the AOGCMs used in Chapter 10. The models are generally run in 'offline' mode, i.e., forced by atmospheric fields derived from high-resolution timeslice experiments, although Huybrechts et al. (2002) and Fichefet et al. (2003) report early efforts at coupling ice sheet models into AOGCMs. Ice sheet models are also included in some EMICs (e.g., Calov et al., 2002). Ridley et al., (2005) point out that the timescale of projected melting of the Greenland ice sheet may be different in coupled and offline simulations. Presently available thermomechanical ice sheet models do not include processes associated with ice streams or grounding-line migration, which may permit rapid dynamical changes in the ice sheets. Glaciers and ice caps, due to their relatively small scales and low likelihood of significant climate feedback on large scales, are not currently included interactively in any AOGCMs. See Chapters 4 and 10 for further detail. For a discussion of terrestrial snow, see Section 8.3.4.1.

8.2.4.2 Sea-Ice

Sea-ice components of current AOGCMs usually predict ice thickness (or volume), fractional cover, snow depth, surface and internal temperatures (or energy), and horizontal velocity. Some models now include prognostic sea ice salinity (Schmidt et al., 2004). Sea ice albedo is typically prescribed with only crude dependence on ice thickness, snow cover and puddling effects.

Since TAR, most AOGCMs have started to employ complex sea ice dynamic components. Complexity of sea-ice dynamics of current AOGCMs vary from the relatively simple "cavitating fluid" model (Flato and Hibler, 1992) to the viscous-plastic model (Hibler, 1979), which is computationally expensive, particularly for global climate simulations. The elastic-viscous-plastic model (Hunke and Dukowicz, 1997) is being increasingly employed, particularly due to its efficiency for parallel computers. New numerical approaches for solving the ice dynamics equations include more accurate representations on curvilinear model grids (Hunke and Dukowicz, 2002; Marsland et al., 2003; Zhang and Rothrock, 2003) and Lagrangian methods for solving the viscous-plastic equations (Lindsay and Stern, 2004; Wang and Ikeda, 2004).

Treatment of sea-ice thermodynamics in AOGCMs has progressed more slowly: typically it includes constant conductivity and heat capacities for ice and snow (if represented), a heat reservoir simulating the effect of brine pockets in the ice, and several layers, the upper one representing snow. More sophisticated thermodynamic schemes are being developed, such as the model of Bitz and Lipscomb (1999), which introduces salinity-dependent conductivity and heat capacities, modeling brine pockets in an energy-conserving way as part of a variable- layer thermodynamic model (e.g., Saenko et al., 2002). Some AOGCMs do include snow-ice formation, which occurs when an ice floe is submerged by the weight of the overlying snow cover and the flooded snow layer refreezes. The latter process is particularly important in the Antarctic sea ice system.

Even with fine grid scales, many sea ice models incorporate sub-grid-scale ice thickness distributions (Thorndike et al., 1975), with several thickness "categories," rather than considering the ice as a uniform slab with inclusions of open water. An ice thickness distribution enables more accurate simulation of thermodynamic variations in growth and melt rates within a single grid cell, which can have significant consequences for ice-ocean albedo feedback processes (e.g., Bitz et al., 2001; Zhang and Rothrock, 2001). A well resolved ice thickness distribution enables a more physical formulation for ice ridging and rafting events, based on energetic principles. Although parameterizations of ridging mechanics and their relationship with the ice thickness distribution have improved (Babko et al., 2002; Toyota et al., 2004; Amundrud et al., 2004), inclusion of advanced ridging parameterizations has lagged other aspects of sea ice dynamics (rheology, in particular) in AOGCMs. Better numerical algorithms used for the ice thickness distribution (Lipscomb, 2001) and ice strength (Hutchings et al., 2004) have also been developed for AOGCMs.

8.2.5 Aerosol Modelling and Atmospheric Chemistry

Climate simulations including atmospheric aerosols with chemical transport have greatly improved since the TAR. Simulated global aerosol distributions are better compared with observations, especially satellite data (e.g., AVHRR, MODIS, MISR, POLDER, TOMS), the ground-based network (AERONET), and many measurement campaigns. (e.g., Chin et al., 2002; Takemura et al., 2002). The global aerosol model inter-comparison project, AEROCOM, has been also initiated in order to improve our understanding of uncertainties of model estimates, and to reduce them (Kinne et al., 2003). These comparisons, combined with cloud observations, should result in improved confidence in the estimation of the aerosol direct and indirect radiative forcing (e.g., Ghan et al., 2001a, 2001b; Lohmann and Lesins, 2002; Takemura et al., 2005). Interactive aerosol subcomponent models have been incorporated in some of the climate models used in Chapter 10 (HadGEM1 and MIROC). Some models also include indirect aerosol effects (e.g. Takemura et al., 2005); however the formulation of these processes is still the subject of much research.

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Interactive atmospheric chemistry components are not generally included in the models used in this report. However CCSM3 includes the modification of greenhouse gas concentrations by chemical processes and conversion of SO_2 and DMS to sulphur aerosols.

8.2.6 Coupling Advances

In an advance since the TAR, a number of groups have developed software allowing easier coupling of the various components of a climate model (e.g., Valcke et al., 2006). An example, the OASIS coupler, developed at CERFACS (Terray et al., 1998), has been used by many modeling centers to synchronize the different models and for the interpolation of the coupling fields between the atmosphere and ocean grids. The schemes for interpolation between the ocean and the atmosphere grids have been revised. The new schemes ensure both a global and local conservation of the various fluxes at the air-sea interface, and track terrestrial, ocean and sea-ice fluxes individually.

Coupling frequency is an important issue, because fluxes are averaged during a coupling interval. Typically most AOGCMs evaluated here, pass fluxes and other variables between the component parts, once per day. The KPP ocean vertical scheme (Large et al., 1994), used in several models, is very sensitive to the wind energy available for mixing. If the models are coupled at a frequency lower than once per ocean timestep, nonlinear quantities such as wind mixing power (which depends on the cube of the wind speed) must be accumulated over every timestep before passing to the ocean. Improper averaging therefore could lead to too little mixing energy and hence shallower mixed layer depths, assuming the parameterization is not re-tuned. However, high coupling frequency can bring new technical issues; in the MIROC model, the coupling interval is 3 hours. In this case, a poorly resolved internal gravity wave is excited in the ocean, and so some smoothing is necessary to damp this numerical problem. It should also be noted that AOGCMs used here, have relatively thick top oceanic grid boxes (typically 10 m or more), limiting the SST response to frequent coupling (Bernie et al. 2005).

8.2.7 Flux Adjustments and Initialization

Since the TAR, more climate models have been developed that do not adjust the surface heat, water and momentum fluxes artificially to maintain a stable control climate. As noted by Stouffer and Dixon (1998), the use of such flux adjustments required relatively long integrations of the component models before coupling. In these models, normally the initial conditions for the coupled integrations were obtained from long spinups of the component models.

In AOGCMs that do not use flux adjustments (see Table 8.1), the initialization methods tend to be more varied. Many models initialize their oceanic components using values obtained either directly from an observationally based, gridded data set (Levitus and Boyer, 1994; Levitus and Antonov, 1997; Levitus et al., 1998) or from short ocean-only integrations that used an observational analysis for their initial conditions. The initial atmospheric component data are usually obtained from atmosphere-only integrations using prescribed SSTs.

To obtain initial data for the preindustrial control integrations discussed in Chapter 10, most AOGCMs use variants of the Stouffer et al. (2004) scheme. In this scheme, the coupled model is initialized as discussed above. The radiative forcing is then set back to preindustrial conditions. The model is integrated for a few centuries using constant preindustrial radiative forcing, allowing the coupled system to partially adjust to this forcing. The degree of equilibration in the real preindustrial climate to the preindustrial radiative forcing is not known. Therefore it seems unnecessary to have the preindustrial control fully equilibrated. After this spin-up integration, the preindustrial control is started and perturbation integrations can begin. An important next step, once the start of the control integration is determined, is the assessment of the control integration climate drift. Large climate drifts can distort both the natural variability (e.g., Inness et al., 2003) and the climate response to changes in radiative forcing (Spelman and Manabe, 1984).

In earlier IPCC reports, the initialisation methods were quite varied. In some cases, the perturbation integrations were initialized using data from control integrations where the SSTs were near present day values and not preindustrial. Given that many climate models now use some variant of the Stouffer et al. method, this situation has improved.

8.3 Evaluation of Contemporary Climate as Simulated by Coupled Global Models

Due to nonlinearities in the processes governing climate, the climate system response to perturbations depends to some extent on its basic state (Spelman and Manabe, 1984). Consequently, for models to predict future climatic conditions reliably, they must simulate the current climatic state with some as yet unknown degree of fidelity. Poor model skill in simulating present climate could indicate that certain physical or dynamical processes have been misrepresented. The better a model simulates the complex spatial patterns and seasonal and diurnal cycles of present climate, the more confidence we can have that all the important processes have been adequately represented. Thus, when new models are constructed, considerable effort is devoted to evaluating their ability to simulate today's climate (e.g., Collins et al., 2006; Delworth et al., 2006).

Some of the assessment of model performance presented here is based on the 20th Century simulations which constitute a part of the multi-model dataset (MMD) archived at PCMDI. In these simulations, modeling groups initiated the models (ca. 1860) from pre-industrial "control" simulations and then imposed the natural and anthropogenic forcing thought to be important for simulating climate of the last 140 years, or so. The twenty-three models considered here (see Table 8.1) are those relied on in Chapters 9 and 10 to investigate historical and future climate changes. Some figures in this section are based on results from a subset of the models because the dataset is incomplete.

In order to identify errors that are systematic across models, the mean of fields available in the MMD, referred to here as the "multi-model mean field," will often be shown. The multi-model mean field results are augmented by results from individual models available as supplementary material (See Figures S8.1 to S8.15)[1]. The multi-model averaging serves to filter out biases of individual models and only retains errors that are generally pervasive. There is some evidence that the multi-model mean field is often in better agreement with observations than any of the fields simulated by the individual models (see Section 8.3.1.1.2), which supports continued reliance on a diversity of modeling approaches in projecting future climate change and provides some further interest in evaluating the multi-model mean results.

Faced with the rich variety of climate characteristics that could potentially be evaluated here, we focus on those elements that can critically affect societies and natural ecosystems and that are most likely to respond to changes in radiative forcing.

8.3.1 Atmosphere

8.3.1.1 Surface Temperature and the Climate System's Energy Budget

For models to simulate accurately the global distribution of the annual cycle and the diurnal cycle of surface temperature, they must, in the absence of compensating errors, correctly represent a variety of processes. The

[1] Supplementary material is available at the website serving the chapter drafts.

large-scale distribution of annual mean surface temperature is largely determined by the distribution of insolation, which is moderated by clouds, other surface heat fluxes, and transport of energy by the atmosphere and to a lesser extent by the ocean. Similarly, the annual and diurnal cycles of surface temperature are governed by seasonal and diurnal changes in these factors, respectively, but they are also damped by storage of energy in the upper layers of the ocean and to a lesser degree the surface soil layers.

8.3.1.1.1 Temperature

Figure 8.2a shows the observed time mean surface temperature as a composite of surface air temperature over regions of land and sea surface temperature (SST) elsewhere. Also shown is the difference between the multi-model mean field and the observed field. With few exceptions, the absolute error (outside polar regions and other data-poor regions) is less than 2 K. Individual models typically have larger errors, but in most cases still less than 3 K, except at high latitudes (see Figure 8.2b and Supplementary Material, Figure S8.1). Some of the larger errors occur in regions of sharp elevation changes and may result simply from mismatches between the model topography (typically smoothed) and the actual topography. There is also a tendency for a slight, but general, cold bias. Outside the polar regions, relatively large errors are evident in the eastern parts of the tropical ocean basins, a likely symptom of problems in the simulation of low clouds. The extent to which these systematic model errors affect a model's response to external perturbations is unknown, but may be significant (see Section 8.6).

[INSERT FIGURE 8.2 HERE]

In spite of the discrepancies discussed here, the fact is that models account for a very large fraction of the global temperature pattern: the pattern correlation between the simulated and observed annual mean temperature is typically about 0.98 for individual models. This supports the view that major processes governing surface temperature climatology are represented with a reasonable degree of fidelity by the models.

An additional opportunity for evaluating models is afforded by the observed annual cycle of surface temperature. Figure 8.3 shows the standard deviation of monthly mean surface temperatures, which is dominated by contributions from the amplitudes of the annual and semi-annual components of the annual cycle. The difference between the mean of the model results and the observations is also shown. The absolute differences are in most regions less than 1 K. Even over extensive land areas of the Northern Hemisphere where the standard deviation generally exceeds 10 K, the models agree with observations within 2 K almost everywhere. The models, as a group, clearly capture the differences between marine and continental environments and also the larger magnitude of the annual cycle found in higher latitudes, but there is a general tendency to underestimate the annual temperature range over eastern Siberia. In general, the largest fractional errors are found over the oceans (e.g., over much of tropical South America and off the east coasts of North America and Asia). These exceptions to the overall good agreement illustrate a general characteristic of current climate models: the largest-scale features of climate are simulated more accurately than regional and smaller scale features.

[INSERT FIGURE 8.3 HERE]

Like the annual range of temperature, the diurnal range (the difference between daily maximum and minimum surface air temperature) is much smaller over oceans than over land (and also better observed), so the discussion here is restricted to continental regions. The diurnal temperature range, zonally and annually averaged over the continents, is generally too small in the models, in many regions by as much as 50% (see Supplementary Material, Figure S8.3). Nevertheless the models simulate the general pattern of this field, with relatively high values over the clearer, drier regions. It is not yet known why models generally underestimate the diurnal temperature range; it is possible that in some models it is in part due to shortcomings of the boundary layer parameterizations or in the simulation of freezing and thawing soil, and it is also known that the diurnal cycle of convective cloud, which interacts strongly with surface temperature, is rather poorly simulated.

Surface temperature is strongly coupled with the atmosphere above it. This is especially evident in mid-latitudes, where migrating cold fronts and warm fronts can cause relatively large swings in surface temperature. Given the strong interactions between the surface temperature and the temperature of the air above, it is of special interest to evaluate how well models simulate the vertical profile of atmospheric temperature. The multi-model mean absolute error in the zonal-mean, annual mean air temperature is almost everywhere less than 2 K (compared with the observed range of temperatures, which spans more than 100 K when the entire troposphere is considered; see Supplementary Material, Figure S8.4). It is notable, however, that near the tropopause at high latitudes the models are generally biased cold. This bias is a problem that has persisted for many years, but in general is now less severe than in earlier models. In a few of the models, the bias has been eliminated entirely, but compensating errors may be responsible. It is known that the tropopause cold bias is sensitive to several factors, including horizontal and vertical resolution, non-conservation of moist entropy, and the treatment of sub-grid scale vertical convergence of momentum ("gravity wave drag"). Although the impact of the tropopause temperature bias on the model's response to radiative forcing changes has not been definitively quantified, it is almost certainly small, relative to other uncertainties.

8.3.1.1.2 The balance of radiation at the top of the atmosphere

The primary driver of latitudinal and seasonal variations in temperature is the seasonally varying pattern of incident sunlight, and the fundamental driver of the circulation of the atmosphere and ocean is the local imbalance between the shortwave (SW) and longwave (LW) radiation at the top of the atmosphere. The impact on temperature of the distribution of insolation can be strongly modified by the distribution of clouds and surface characteristics.

Considering first the annual mean shortwave flux at the "top" of the atmosphere (TOA)[2], the insolation is determined by well-known orbital parameters that ensure good agreement between models and observations. The annual mean insolation is strongest in the tropics, decreasing to about half as much at the poles. This largely drives the strong equator to pole temperature gradient. As for outgoing SW, the Earth, on average, reflects about the same amount of sunlight (~100 W m^{-2}, in the annual mean) at all latitudes. At most latitudes, the difference between the multi-model mean zonally averaged outgoing SW and observations is in the annual mean less than 6 W m^{-2} (i.e., an error of about 6%; see Supplementary Material, Figure S8.5). Given that clouds are responsible for about half the outgoing SW, these errors are not surprising, for it is known that cloud processes are among the most difficult to simulate by models (see Section 8.6.3.2.3).

There are additional errors in outgoing SW radiation due to variations with longitude and season, and these can be quantified by means of the root-mean-square (RMS) error, calculated for each latitude over all longitudes and months and plotted in Figure 8.4a (see also Supplementary Material, Figure S8.6). Errors in the complete two-dimensional fields (see Supplementary Material, Figure S8.6) tend to be substantially larger than the zonal mean errors of about 6 W m^{-2}, an example of the common result that model errors tend to increase as smaller spatial scales and shorter time scales are considered. Figure 8.4a also illustrates a common result that the errors in the multi-model average of monthly mean fields are often smaller than the errors in the individual model fields. In the case of outgoing SW radiation, this is true at nearly all latitudes. Calculation of the global mean RMS error, based on the monthly mean fields and area-weighted over all grid cells, indicates that the individual model errors are in the range 15–22 W m^{-2}, whereas the error in the multi-model mean climatology is only 13.1 W m^{-2}. Why the multi-model mean field turns out to be closer to the observed than the fields in any of the individual models is the subject of ongoing research; a superficial explanation is that at each location and for each month, the model estimates tend to scatter around the correct value (more or less symmetrically), with no single model consistently closest to the observations. This, however, does not explain *why* the results should scatter in this way.

[INSERT FIGURE 8.4 HERE]

At the top of the atmosphere the net shortwave radiation is everywhere partially compensated by outgoing LW radiation (i.e., infrared emissions) emanating from the surface and the atmosphere. Globally and annually averaged, this compensation is nearly exact. The pattern of LW radiation emitted by earth to space depends most critically on atmospheric temperature, humidity, clouds, and surface temperature. With a few exceptions, the models can simulate the observed zonal mean of the annual mean outgoing LW within 10 W m^{-2} (an error of around 5%; see Supplementary Material, Figure S8.7). The models reproduce the relative

[2] The atmosphere clearly has no identifiable "top", but the term is used here to refer to an altitude above which the absorbtion of shortwave and longwave radiation is negligibly small.

minimum in this field near the equator where the relatively high humidity and extensive cloud cover in the tropics raises the effective height (and lowers the effective temperature) at which LW radiation emanates to space.

The seasonal cycle of the outgoing LW radiation pattern is also reasonably well simulated by models (see Figure 8.4b). The RMS error for most individual models varies from about 3% of the OLR near the poles to somewhat less than 10% in the tropics. The errors for the multi-model mean simulation, ranging from about 2% to 6% across all latitudes, are again generally smaller than those in the individual models.

For a climate in equilibrium, any local annual mean imbalance in the net TOA radiative flux (SW + LW) must be balanced by a vertically integrated net horizontal divergence of energy carried by the ocean and atmosphere. The fact that the TOA SW and LW fluxes are well simulated implies that the models must also be properly accounting for poleward transport of total energy by the atmosphere and ocean. This proves to be the case with most models correctly simulating poleward energy transport within about 10%. Although superficially this would seem to provide an important check on models, it is likely that in current models compensating errors improve their agreement with observations. There are in fact theoretical and model studies that suggest that if the atmosphere fails to transport the observed portion of energy, the ocean will tend to largely compensate (e.g., Shaffrey and Sutton, 2004).

8.3.1.2 Moisture and Precipitation

Water is fundamental to life, and if regional, seasonal precipitation patterns were to change, the potential impacts could be profound. Consequently, it is of real practical interest to evaluate how well models can simulate precipitation, not only at global scales, but also regionally. Unlike seasonal variation of the temperature, which at large scales is strongly determined by the insolation pattern and configuration of the continents, precipitation variations also are strongly influenced by vertical movement of air due to atmospheric instabilities of various kinds and by the flow of air over orographic features. For models to simulate accurately the seasonally varying pattern of precipitation, they must correctly simulate a number of processes (e.g., evapo-transpiration, condensation, transport) that are difficult to evaluate on a global scale. Some of these are discussed further in Sections 8.2 and 8.6. Here the focus will be on the distribution of precipitation and water vapor.

Figure 8.5a shows observation-based estimates of annual mean precipitation and Figure 8.5b shows the multi-model mean field. At the largest scales, the lower precipitation rates at higher latitudes reflect both reduced local evaporation at lower temperatures and a lower saturation vapor pressure of cooler air, which tends to inhibit the transport of vapor from other regions. In addition to this large-scale pattern, captured well by models, is a local minimum in precipitation near the equator in the Pacific, due to a tendency for the intertropical convergence zone (ITCZ[3]) to reside off the equator. There are local maxima in mid-latitudes, reflecting the tendency for subsidence to suppress precipitation in the subtropics and for storm systems to enhance precipitation in mid-latitudes. The models capture these large-scale zonal mean precipitation differences, suggesting that they can adequately represent these features of atmospheric circulation. Moreover there is some evidence provided in Section 8.3.5 that models have improved over the last several years in simulating the annual cycle of the precipitation patterns.

[INSERT FIGURE 8.5 HERE]

Models also simulate some of the major regional characteristics of the precipitation field, including the major convergence zones and the maxima over tropical rain forests, although there is a tendency to underestimate rainfall over the Amazon. When considered in more detail, however, there are also deficiencies in the multi-model mean precipitation field. There is a distinct tendency for models to orient the South Pacific convergence zone parallel to latitudes and to extend it too far eastward. In the tropical Atlantic the precipitation maximum is too weak in most models with too much rain south of the equator. There are also systematic east-west positional errors in the precipitation distribution over the Indo-Pacific Warm Pool

[3] The ITCZ is manifested as a band of relatively intense convective precipitation, accompanied by surface convergence of moisture, whch tends to locate seasonally over the warmest surface temperatures and circumnavigates the earth in the tropics (though not continuously).

in most models, with an excess of precipitation over the Western Indian Ocean and over the Maritime Continent. These lead to systematic biases in the location of the major rising branches of the Walker Circulation and can compromise major teleconnection[4] pathways, in particular those associated with El Niño (e.g., Turner et al., 2005). Systematic dry biases over the Bay of Bengal are related to errors in the monsoon simulations.

Despite the apparent skill suggested by the multi-model mean (Figure 8.5), many models, individually, display substantial precipitation biases, especially in the tropics, which often approach the magnitude of mean observed climatology (e.g., Johns et al. 2006; also see Supplementary Material, Figures S8.9, S8.10). Although some of these biases can be attributed to errors in the SST field of the coupled model, even atmosphere-only versions of the models show similarly large errors (e.g., Slingo et al. 2003). This may be one factor leading to a lack of consensus among models even as to the sign of future regional precipitation changes predicted in parts of the tropics (see Chapter 10).

At the heart of understanding what determines the regional distribution of precipitation over land and oceans in the tropics is atmospheric convection and its interaction with large-scale circulation. Convection occurs on a wide range of space and time scales, and there is increasing evidence that interactions across all scales may be crucial for determining the mean tropical climate and its regional rainfall distributions (e.g., Khairoutdinov et al., 2005). Over tropical land, the diurnal cycle dominates, and yet many models have difficulty in simulating the early evening maximum in rainfall. Instead, they systematically tend to rain before noon (Yang and Slingo, 2001; Dai, 2006), which compromises the energy budget of the land surface. Similarly the land-sea breezes around the complex system of islands in Indonesia have been implicated in the failure of models to capture the regional rainfall patterns across the Indo-Pacific Warm Pool (Neale and Slingo, 2003). Over the oceans, the precipitation distribution along the ITCZ results from organised convection associated with weather systems occurring on synoptic and intraseasonal timescales (e.g., MJO – See Section 8.4.8). These systems are frequently linked to convectively-coupled equatorial wave structures (e.g. Yang et al. 2003), but these are poorly represented in models (e.g., Ringer et al., 2006; Lin et al., 2006). Thus the rain-bearing systems, which establish the mean precipitation climatology, are not well simulated, contributing also to the poor temporal characteristics of daily rainfall (e.g., Dai, 2006), in which many models rain too frequently but with reduced intensity.

Precipitation patterns are intimately linked to atmospheric humidity, evaporation, condensation and transport processes. Good observational estimates of the global pattern of evaporation are not available, and condensation and vertical transport of water vapor can often be dominated by subgrid scale convective processes which are difficult to evaluate globally. The best prospect for assessing water vapor transport processes in humid regions, especially at annual and longer time scales, may be to compare modeled and observed stream flow which must nearly balance atmospheric transport, since terrestrial water storage variations on longer timescales are small (Milly et al., 2005; see Section 8.3.4.2).

Although an analysis of runoff in the multi-model dataset at PCMDI has not yet been performed, the net result of evaporation, transport and condensation processes can be seen in the atmospheric humidity distribution. Models reproduce the large-scale decrease of humidity with both latitude and altitude (see Supplementary Material, Figure S8.11), although this is not truly an independent check of models, since it is almost a direct consequence of their reasonably realistic simulation of temperature. The multi-model mean bias in humidity, zonally and annually averaged, is less than 10% throughout most of the lower troposphere compared with reanalyses, but model evaluation in the upper troposphere is considerably hampered by observational uncertainty.

Any errors in the water vapor distribution should impact the outgoing LW radiation (see Section 8.3.1.1.2), which was seen to be free of systematic zonal mean biases. In fact, the observed differences in outgoing LW radiation between the moist and dry regions are reproduced by the models, providing some confidence that any errors in humidity are not critically impacting the net fluxes at the top of the atmosphere. The strength of water vapor feedback, which strongly affects global climate sensitivity, is, however, primarily determined by

[4] Teleconnection describes the process through which changes in one part of the climate system affect a remote location via changes in atmospheric ciculation patterns.

fractional *changes* in water vapor in response to warming, and the ability of models to correctly represent this feedback is perhaps better assessed with process studies (see Section 8.6).

8.3.1.3 Extra-Tropical Storms

The impact of extra-tropical cyclones on global climate derives primarily from their role in transporting heat, momentum and humidity. Regionally and individually these mid-latitude storms often provide beneficial precipitation, but also occasionally produce destructive flooding and high winds. For these reasons the effect of climate change on extra-tropical cyclones is of considerable importance and interest.

Among the several approaches used to characterize cyclone activity (e.g., Paciorek et al., 2002), analysis methods that identify and track extra-tropical cyclones can provide the most direct information concerning their frequency and movement (Hoskins and Hodges, 2002, 2005). Climatologies for the distribution and properties of cyclones found in models can be compared with reanalysis products (Chapter 3), which provide the best observation-constrained data.

Results from a systematic analysis of AMIP-2 simulations (PCMDI, 2004) indicate that models run with observed sea surface temperatures are capable of producing storm tracks located in about the right locations, but nearly all show some deficiency in the distribution and level of cyclone activity. In particular, simulated storm tracks are often more zonally oriented than is observed. A study by Lambert and Fyfe (2006), based on the multi-model dataset at PCMDI, finds that as a group, the recent models, which include interactive oceans, tend to underestimate slightly the total number of cyclones in both hemispheres. The number of *intense* storms, however, is slightly overestimated in the Northern Hemisphere, but underestimated in the Southern Hemisphere, although observations are less certain there.

Increases in model resolution (characteristic of models over the last several years) appear to improve some aspects of extra-tropical cyclone climatology (Bengtsson et al., 2006), particularly in the Northern Hemisphere where observations are most reliable (Hodges et al., 2003; Hanson et al., 2004; Wang et al., 2006). Improvements to the dynamical core and physics of models have also led to better agreement with reanalyses (Ringer et al., 2006; Watterson, 2006)

Our assessment is that although problems remain, climate models are improving in their simulation of extra-tropical cyclones.

8.3.2 Ocean

As noted earlier, we focus only on those variables important in determining the transient response of climate models (see Section 8.6). Due to space limitations, much of the analysis performed for this section is found in the supplementary material (Figures S8.12 to S8.15). An assessment of the modes of natural, internally generated variability can be found in Section 8.4. Comparisons of the type performed here need to be made with an appreciation of the uncertainties in the historical estimates of radiative forcing and various sampling issues in the observations (see Chapters 2, 5). Unless otherwise noted, all results discussed here are based on the multi-model dataset at PCMDI.

8.3.2.1 Simulation of Mean Temperature and Salinity Structure

Before discussing the oceanic variables directly involved in determining the climatic response, it is important to discuss the fluxes exchanged between the ocean and atmosphere. Modelling experience shows that the surface fluxes play a large part in determining the fidelity of the oceanic simulation. Of course, the atmosphere and ocean are coupled, so that the fidelity of the oceanic simulation feeds back on the atmospheric simulation, impacting the surface fluxes.

Unfortunately the total surface heat and water fluxes (see Supplementary Material, Figure S8.14) are not well observed. Normally, they are inferred from observations of other fields, such as surface temperature and winds. Consequently, the uncertainty in the observational estimate is large – of the order of tens of W m^{-2} for the heat flux, even in the zonal mean. An alternative way of assessing the surface fluxes is by looking at the horizontal transports in the ocean. In a long term average, the heat and water storage in the ocean is small so that the horizontal transports have to balance the surface fluxes. Since the heat transport seems better constrained by the available observations, it is presented here.

North of 45°N, most models transport too much heat northward when compared to the observational estimates used here (Figure 8.6), but there is uncertainty in the observations. At 45° N, for example, the models lie much closer to the estimate of 0.6 PW obtained by Ganachaud and Wunsch (2003). From 45°N to the equator, most model estimates lie near or between the observational estimates shown. In the tropics and subtropical zone of the Southern Hemisphere, most models underestimate the southward heat transport away from the equator. In middle and high latitudes of the Southern Hemisphere, the observational estimates are more uncertain and the model heat transports tend to surround the observational estimates.

[INSERT FIGURE 8.6 HERE]

The oceanic heat fluxes have large seasonal variations which lead to large variations in the seasonal storage of heat by the oceans, especially in mid-latitudes. The oceanic heat storage tends to damp and delay the seasonal cycle of surface temperature. The models evaluated here agree well with the observations of seasonal heat storage by the oceans (Gleckler et al., 2006a). The most notable problem area for the models is in the tropics, where many models continue to have biases in representing the flow of heat from the tropics into middle and high latitudes.

The annually averaged, zonal component of surface wind stress, zonally averaged over the oceans, is reasonably well simulated by the models (Figure 8.7). At most latitudes, the reanalysis estimates (based on atmospheric models constrained by observations) lie within the range of model results. In middle to low latitudes, the model spread is relatively small and all the model results lie fairly close to the reanalysis. In middle to high latitudes, the model simulated wind stress maximum tends to lie equatorward of the reanalysis. This error is particularly large in the Southern Hemisphere, a region where there is more uncertainty in the reanalysis. Almost all models place the Southern Hemisphere wind stress maximum north of the reanalysis estimate. The Southern Ocean wind stress errors in the control integrations may adversely impact other aspects of the simulation and possibly the oceanic heat uptake under climate change, as discussed below.

[INSERT FIGURE 8.7 HERE]

The largest individual model errors in the zonally averaged sea surface temperature (SST) (Figure 8.8) are found in middle and high latitudes, particularly in the middle latitudes of the Northern Hemisphere where the model temperatures are too cold. Almost every model has some tendency for this cold bias. This error seems to be associated with the poor simulation of the path of the North Atlantic Current and seems to be due to an ocean component problem rather than a problem with the surface fluxes. In the zonal averages near 60°S, there is a warm bias in the multi-model mean results. Many models suffer from a too warm bias in the Southern Ocean SSTs.

[INSERT FIGURE 8.8 HERE]

In the individual model SST error maps (see Supplementary Material, Figure S8.1), it is apparent that most models have a large warm bias in the eastern parts of the tropical ocean basins, near the continental boundaries. This is also evident in the multi-model mean result (Figure 8.2a) and is associated with insufficient resolution which leads to problems in the simulation of the local wind stress, oceanic upwelling and under-prediction of the low cloud amounts (see Sections 8.2 and 8.3.1). These are also regions where there is a relatively large spread among the model simulations indicating a relatively wide range in the magnitude of these errors. Another area where the model error spread is relatively large is found in the North Atlantic Ocean. As noted above, this is an area where many models have problems properly locating the North Atlantic Current, a region of large SST gradients.

In spite of the errors, the model simulation of the SST field is fairly realistic overall. Over all latitudes, the multi-model mean zonally averaged SST error is less than 2 K, which is fairly small considering that most models do not use flux adjustments in these simulations. The model mean local SST errors are also less than

2 K over most regions, with only relatively small areas exceeding this value. Even relatively small SST errors, however, can adversely impact the simulation of variability and teleconnections (Section 8.4).

Over most latitudes, at depths ranging from 200 to 3000 m, the multi-model mean zonally averaged ocean temperature is too warm (see Figure 8.9). The maximum warm bias (about 2 K) is located in the region of the North Atlantic Deep Water (NADW) formation. Above 200 m, however, the multi-model mean is too cold, with maximum cold bias (more than 1 K) near the surface in mid-latitudes of the Northern Hemisphere, as discussed above. Most models generally have an error pattern similar to the multi-model mean (see Supplementary Material, Figure S8.12) except for CNRM-CM3 and MRI-CGCM2.3.2 which are too cold throughout most of the middle and low latitude ocean (see Supplementary Material, Figure S8.12). The GISS-EH model is much too cold throughout the subtropical thermocline and only the Northern Hemisphere part of the FGOALS-g1.0 error pattern is similar to the model mean error described here. The magnitude of these errors, especially in the deeper parts of the ocean, depends on the AOGCM initialization method (Section 8.2.7).

[INSERT FIGURE 8.9 HERE]

The error pattern in which the upper 200 meters of the ocean tend to be too cold, while the layers below are too warm indicates that the thermocline in the multi-model mean is too diffuse. This error, which was also present at the time of the TAR, seems partly related to the wind stress errors in the Southern Hemisphere noted above and possibly to errors in formation and mixing of North Atlantic Deep Water. The multi-model mean errors in temperature (too warm) and salinity (too salty; see Supplementary Material, Figure S8.13) in middle and low latitudes near the base of the thermocline tend to cancel in terms of a density error and appear to be associated with the problems in the formation of Antarctic Intermediate Water (AAIW), as discussed below.

8.3.2.2 Simulation of Circulation Features Important for Climate Response

8.3.2.2.1 Meridional overturning circulation

The meridional overturning circulation (MOC) is an important component of present day climate and many models indicate that it will change in the future (Chapter 10). Unfortunately many aspects of this circulation are not well observed. The MOC transports large amounts of heat and salt into high latitudes of the North Atlantic Ocean. There the relatively warm, salty surface waters are cooled by the atmosphere, making the water dense enough to sink to depth. These waters then flow southward towards the Southern Ocean where they mix with the rest of the World Ocean waters (see Supplementary Material, Figure S8.15).

The models simulate this major aspect of the MOC and also simulate a number of distinct wind driven surface cells (see Supplementary Material, Figure S8.15). In the tropics and subtropics, these cells are quite shallow, but at the latitude of the Drake Passage (55°S), the wind-driven cell extends to a much greater depth (2 to 3 km). Most models in the multi-model dataset have some manifestation of the wind driven cells. The strength and pattern of the overturning circulation varies greatly from model to model (see Supplementary Material, Figure S8.15). GISS-AOM exhibits the strongest overturning circulation, with almost 40 to 50 Sv (10^6 m^3/s). The CGCM (T47 and T63), FGOALS have the weakest overturning circulations, about 10 Sv. The observed value is about 18 Sv (Ganachaud and Wunsch 2000).

In the Atlantic, the overturning circulation, extending to considerable depth, is responsible for a large fraction of the northward oceanic heat transport, in both observations and models (e.g., Hall and Bryden, 1982; Gordon et al., 2000). Figure 10.15 contains an index of the Atlantic MOC at 30°N for the suite of AOGCM 20th Century simulations. While the majority of models show an MOC strength that is within observational uncertainty, some show higher and lower values and a few show substantial drifts which could make interpretation of MOC projections using those models very difficult.

Overall, some aspects of the simulation of the MOC have improved since the TAR. This is due in part to improvements in mixing schemes, the use of higher resolution ocean models (see Section 8.2), and better simulation of the surface fluxes. This improvement can be seen in the individual model MOC sections (see Supplementary Material, Figure S8.15) by the fact that (1) the location of the deep water formation is more realistic, with more sinking occurring in the Greenland-Iceland-Norwegian (GIN) and Labrador Seas as evidenced by the larger streamfunction values north of the sill located at 60°N (e.g., Wood et al., 1999) and (2) deep waters are subjected to less spurious mixing, resulting in better water mass properties (Thorpe et al., 2004) and a larger fraction of the water that sinks in the northern part of the North Atlantic Ocean exiting the Atlantic Ocean near 30S (Danabasoglu et al., 1995). There is still room for improvement in the models' simulation of these processes, but there is clear evidence of improvement in many of the models analyzed here.

8.3.2.2.2 Southern ocean circulation

The Southern Ocean wind stress error has a particularly large detrimental impact on the Southern Ocean simulation in the models. Partly due to the wind stress error identified above, the location of the Antarctic Circumpolar Current (ACC) is also too far north in most models (Russell et al., 2006). Since the AAIW is formed on the north side of the ACC, the water mass properties of the AAIW are distorted (typically too warm and salty - Russell et al., 2006). The relatively poor AAIW simulation contributes to the multi-model mean error identified above where the thermocline is too diffuse, because the waters near the base of thermocline are too warm and salty.

It is likely that the relatively poor Southern Ocean simulation will influence the transient climate response to increasing greenhouse gases by impacting the oceanic heat uptake. When forced by increases in radiative forcing, models with too little Southern Ocean mixing will probably underestimate the ocean heat uptake; models with too much mixing will likely exaggerate it. These errors in oceanic heat uptake will also have a large impact on the reliability of the sea level rise projections. See Chapter 10 for more discussion on this subject.

8.3.2.3 Summary of Oceanic Component Simulation

Overall, the improvements in the simulation of the observed time mean ocean state noted in the TAR (McAvaney et al., 2001) have continued in the models evaluated here. It is notable that this improvement has continued in spite of the fact that nearly all models no longer use flux adjustments. This suggests that the improvements in the physical parameterizations, increased resolution (see Section 8.2) and improved surface fluxes are together having a positive impact on model simulations. The temperature and salinity errors in the thermocline, while still large, have been reduced in many models. In the Northern Hemisphere, many models still suffer from a cold bias in the upper ocean which is a maximum near the surface and may distort the ice-albedo feedback in some models (see Section 8.3.3). In the Southern Ocean, the equatorward bias of the westerly wind stress maximum found in most models is a problem that may affect the models’ response to increasing radiative forcing.

8.3.3 Sea Ice

The magnitude and spatial distribution of the high-latitude climate changes can be strongly affected by sea ice characteristics, but evaluation of sea-ice in models is hampered by insufficient observations of some key variables (e.g., ice thickness) (see Section 4.4). Even when sea-ice errors can be quantified, it is difficult to isolate their causes, which might arise from deficiencies in the representation of sea ice itself, but could also be due to flawed simulation of the atmospheric and oceanic fields in high latitudes which drive ice movement (see Sections 8.3.1, 8.3.2, 11.3.8).

Although sea ice treatment in AOGCMs has become more sophisticated, including better representation of both the dynamics and thermodynamics (see Section 8.2.4), improvement in simulating sea ice in these models, as a group, is not obvious (compare Figure 8.10 with TAR Figure 8.10; or Kattsov and Källén, 2005, Figure 4.11). In some models, however, the geographical distribution and seasonality of sea ice is now better reproduced.

[INSERT FIGURE 8.10 HERE]

For the purposes of model evaluation, the most reliably measured characteristic of sea ice is its seasonally varying extent (i.e., the area enclosed by the ice edge, operationally defined as the 15% contour; see Section 4.4). Despite the wide differences among the models, the multi-model mean of sea ice extent is in a reasonable agreement with observations. Based on fourteen of the fifteen AOGCMs available at the time of analysis (one model was excluded because of unrealistically large ice extents (Arzel et al., 2006)), the mean extent of simulated sea ice exceeds that observed in the Northern Hemisphere (NH) by up to roughly 10^6 km^2 throughout the year, whereas in the Southern Hemisphere (SH) the annual cycle is exaggerated, with too much sea ice in September (~2×10^6 km^2) and too little in March by a lesser amount. In many models the regional distribution of sea ice is poorly simulated, even if the hemispheric areal extent is approximately correct (Arzel et al., 2006; Holland and Raphael, 2006; Zhang and Walsh, 2006). The spread of simulated sea ice extents, measured as the multi-model standard deviation from the model mean, is generally narrower in the NH than in the SH (Arzel et al., 2006). Even in the best case (NH winter), the range of simulated sea ice extent exceeds 50% of the mean, and ice thickness also varies considerably, suggesting that projected decreases in sea ice cover remain rather uncertain. The model sea ice biases may influence global climate sensitivity (see Section 8.6). There is a tendency for models with relatively large sea ice extent in the present climate to have higher sensitivity. This is apparently especially true of models with low to moderate polar amplification (Holland and Bitz, 2003).

Among the primary causes of biases in simulated sea ice (especially its distribution) are biases in the simulation of high latitude winds (Walsh et al., 2002; Chapman and Walsh, 2006; Bitz et al., 2002), as well as vertical and horizontal mixing in the ocean (Arzel et al., 2006). Also important are surface heat flux errors, which in particular may result from inadequate parameterizations of the atmospheric boundary layer (under stable conditions commonly occurring at night and in the wintertime over sea ice) and generally from poor simulation of high latitude cloudiness, which is evident from the striking inter-model scatter (e.g., Kattsov and Källén, 2005).

8.3.4 Land Surface

Our ability to evaluate the land surface component in coupled models is severely limited by the lack of suitable observations. The terrestrial surface plays key climatic roles in influencing the partitioning of available energy between sensible and latent heat fluxes, determining whether water drains or remains available for evaporation, determining the surface albedo and whether snow melts or remains frozen, and influencing surface fluxes of carbon and momentum. Few of these can be evaluated at large spatial or long temporal scales. This section therefore evaluates those quantities for which some observational data exist.

8.3.4.1 Snow Cover

Analysis and comparison of AMIP-2 results, available at the time of the TAR, and more recent AOGCM results in the present multi-model dataset (MMD) at PCMDI show that models are now more consistent in their simulation of snow cover. Problems remain however, and Roesch (2006) showed that the recent models predict excessive snow water equivalent (SWE) in spring, likely because of excessive winter precipitation. Frei et al. (2005) found that AMIP-2 models simulate the seasonal timing and the relative spatial patterns of SWE over North America fairly well. A tendency to overestimate ablation during spring was however identified. On the continental scale, the highest monthly SWE integrated over the North American continent in AMIP-2 models varies within ±50% of the observed value of ~1500 km^3. The magnitude of these model errors is large enough to affect continental water balances. Snow cover area (SCA) in the recent models is well captured, but interannual variability is too low during melt. Frei et al., 2003 showed where observations were within the interquartile range of AMIP-2 models for all months at the hemispheric and continental scale. Encouragingly, there was significant improvement over earlier AMIP-1 simulations for seasonal and interannual variability of SCA (Frei et al., 2005). Both the recent AOGCM's and AMIP models reproduced the observed decline in annual SCA over the period 1979–1995 and most models captured the observed decadal scale variability over the 20th century. Despite these improvements, a minority of models still exaggerate SCA.

Large discrepancies remain in albedo for forested areas under snowy conditions due to difficulties in determining the extent of masking of snow by vegetation (Roesch, 2006). The ability of terrestrial models to simulate snow under observed meteorological forcing has been evaluated via several intercomparisons. At the scale of individual grid cells, for mid-latitude (Slater et al., 2001) and alpine (Etchevers et al., 2004) locations, the spread of model simulations usually encompass observations. However, grid-box scale simulations of snow over high-latitude river basins identified significant limitations (Nijssen et al., 2003), due to difficulties relating to calculating net radiation, fractional snow cover, and interactions with vegetation.

8.3.4.2 Land Hydrology

The evaluation of the hydrological component of climate models has mainly been conducted uncoupled (Bowling et al., 2003; Nijssen et al., 2003; Boone et al., 2004). This is due in part to the difficulties of evaluating runoff simulations across a range of climate models due to variations in rainfall, snow melt and net radiation. Some attempts have, however, been made. Arora (2001) used the AMIP-2 framework to show that the Canadian Climate Model's simulation of the global hydrological cycle compared well to observations, but regional variations in rainfall and runoff led to differences at the basin-scale. Gerten et al. (2004) evaluated the hydrological performance of the Lund-Potsdam-Jena (LPJ) model and showed that the model performed well in the simulation of runoff and evapotranspiration compared to other global hydrological models, although the version of LPJ assessed had been enhanced to improve the simulation of hydrology over the versions used by Sitch et al. (2003).

Milly et al. (2005) made use of the multi-model dataset, which contains results from recent models, to investigate whether observed 20th-Century trends in regional land hydrology could be attributed to variations in atmospheric composition and solar irradiance. Their analysis, based on an ensemble of 26 integrations of 20th Century climate from nine climate models, showed that at regional scales these models simulated observed stream flow measurements with good qualitative skill. Further, the models demonstrated highly significant quantitative skill in identifying the regional runoff trends indicated by 165 long-term stream gauges. They concluded that the impact of changes in atmospheric composition and solar irradiance on observed stream flow was, at least in part, predictable. This is an important scientific advance: it suggests that despite limitations in the hydrological parameterizations included in climate models, these models can capture observed changes in 20th Century stream flow associated with atmospheric composition and solar irradiance changes. This enhances our confidence in the use of these models for future projection.

8.3.4.3 Surface Fluxes

Despite considerable effort since the TAR, uncertainties remain in the representation of solar radiation in climate models (Potter and Cess, 2004). The AMIP-2 results and the recent model results in the MMD provide an opportunity for a major systematic evaluation of model ability to simulate solar radiation. Wild (2005) and Wild et al. (2006) evaluated these models and found considerable differences in the global annual mean solar radiation absorbed at the Earth's surface. In comparison to global surface observations, Wild (2005) concludes that many climate models overestimate surface absorption of solar radiation partly due to problems in the parameterizations of atmospheric absorption, clouds and aerosols. Similar uncertainties exist in the simulation of downwelling infrared radiation (Wild et al., 2001). Difficulties in simulating absorbed solar and infrared radiation at the surface leads inevitable to uncertainty in the simulation of surface sensible and latent heat fluxes.

8.3.4.4 Carbon

A major advance since the TAR is some systematic assessments of the capability of land surface models to simulate carbon. Dargaville et al. (2002) evaluated the capacity of four global vegetation models to simulate the seasonal dynamics and interannual variability of atmospheric CO_2 between 1980 and 1991. Using off-line forcing, they evaluated the capacity of these models to simulate carbon fluxes, via an atmospheric transport model, using observed atmospheric CO_2. They found that the terrestrial models tended to underestimate the amplitude of the seasonal cycle and simulated the spring uptake of CO_2 approximately 1–2 months too early. Of the four models, none was clearly superior in its capacity to simulate the global carbon budget, but all four reproduced the main features of the observed seasonal cycle in atmospheric CO_2. A further off-line evaluation of the LPJ global vegetation model by Sitch et al. (2003) provided confidence that the model could replicate the observed vegetation pattern, seasonal variability in net ecosystem exchange and local soil moisture measurements when forced by observed climatologies.

The only systematic evaluation of carbon models that were interactively coupled to climate models occurred as part of C4MIP where Friedlingstein et al. (2006) compared a suite of models' capacity to simulate historical CO_2 forced by observed emissions. Issues relating to the magnitude of the fertilization effect and the partitioning between land and ocean uptake were identified in individual models, but it is only under increasing CO_2 in the future (see Chapter 10) that the differences become large. Several other groups have evaluated the impact of coupling specific models of carbon into climate models but clear results are difficult to obtain because of inevitable biases in both the terrestrial and atmospheric modules (e.g., Delire et al., 2003).

8.3.5 Changes in Model Performance

Standard experiments, agreed upon by the climate modeling community to facilitate model intercomparison (see Section 8.1.2.2), have produced archives of model output that make it easier to track historical changes in model performance. Most of the modeling groups that contributed output to the current multi-model dataset (MMD) at PCMDI also archived simulations from their earlier models (ca. 2000) as part of the Coupled Model Intercomparison Project (CMIP1&2). The TAR largely relied on the earlier generation of models in its assessment.

Based on the archived model output, it is possible to quantify changes in performance of evolving models. This can be done most straightforwardly by only considering the 14 modeling groups that contributed output from both their earlier and more recent models.[5] One important aspect of model skill is how well the models simulate the seasonally varying global pattern of climatically important fields. The only monthly mean fields available in the CMIP1&2 archive are surface air temperature, precipitation and mean sea-level pressure, so these are the focus of this analysis. Although the simulation conditions in the MMD 20th Century simulations were not identical to those in the CMIP1&2 control runs, the differences do not alter the conclusions summarized below because the large-scale climatological features dominate, not the relatively small perturbations resulting from climate change.

A summary of the ability of AOGCMs to simulate the seasonally varying climate state is provided by Figure 8.11 which displays error measures that gauge how well recent models simulate precipitation, sea-level pressure, and surface temperature, compared with their predecessors. The normalized root-mean-square (RMS) error shown is a so-called space-time statistic, computed from squared errors, summed over all twelve climatological months and over the entire globe, with grid-cell values weighted by the corresponding grid-cell area. This statistic can be used to assess the combined contributions of both spatial pattern errors and seasonal cycle errors. The RMS error is divided by the corresponding observed standard deviation of the field to provide a relative measure of the error. In Figure 8.11 this scaling implies that the pressure is better simulated than precipitation, and that surface temperature is simulated best of all.

[INSERT FIGURE 8.11 HERE]

The models in Figure 8.11 are categorized base on whether or not flux adjustments were applied (see Section 8.2.7). Of the earlier generation models, eight of the fourteen models were flux adjusted, but only two of these groups continue this practice. Several conclusions can be arrived at based on the figure: 1) although flux-adjusted models on average have smaller errors than those without (in both generations), the smallest errors in simulating sea-level pressure and surface temperature are found in models without flux adjustment; 2) despite the elimination of flux adjustment in all but two of the recent models, the mean error obtained from the recent suite of 14 models is smaller than errors found in the corresponding earlier suite of models; and 3) models without flux adjustment have improved on average, as have the flux-adjusted models. An exception to this last statement is the slight increase in mean RMS error for sea-level pressure found in non-flux-adjusted models. Despite no apparent improvement in the mean in this case, three of the recent generation models have smaller sea-level pressure errors than any of the earlier models.

These results demonstrate that the models now being used in applications by major climate modeling groups better simulate seasonally varying patterns of precipitation, mean sea-level pressure and surface air temperature than the models relied on by these same groups at the time of the TAR.

[5] One modeling group participating in CMIP1&2 did not contribute to the MMD, and four groups providing output to the MMD did not do so for CMIP1&2. Results from these five groups are therefore not considered in this subsection. Some modeling groups contributed results from more than one version of their model (sometimes, simply running it at two different resolutions), and in these cases the mean of the two model results is considered here.

8.4 Evaluation of Large-Scale Climate Variability as Simulated by Coupled Global Models

The atmosphere-ocean coupled climate system shows various modes of variability that range widely from intraseasonal to interdecadal time-scales. Successful simulation and prediction over a wide range of these phenomena increases our confidence in the AOGCMs used for climate predictions of the future.

8.4.1 Northern and Southern Annular Modes (NAM and SAM)

There is evidence (e.g., Fyfe et al., 1999; Shindell et al., 1999) that the simulated response to greenhouse gas forcing in AOGCMs has a pattern that resembles the models' NAM, and thus it would appear important that the NAM (see Chapters 3 and 9) is realistically simulated. Analyses of individual AOGCMs (e.g., Fyfe et al., 1999; Shindell et al., 1999) have demonstrated that they are capable of simulating many aspects of the NAM and NAO patterns including linkages between circulation and temperature. Multi-model comparisons (for winter atmospheric pressure, Osborn (2004); for winter temperature, Stephenson and Pavan (2003); and for atmospheric pressure across all months of the year, AchutaRao et al. (2004)), including assessments of the multi-model dataset at PCMDI (Miller et al., 2006) confirm the overall skill of AOGCMs but also identify that teleconnections between the Atlantic and Pacific Oceans are stronger in many models than is observed (Osborn, 2004). In some models this is related to a bias towards a strong polar vortex in all winters so that their simulations nearly always reflect behaviour that is only observed at times with strong vortices (when a stronger Atlantic–Pacific correlation is observed, Castanheira and Graf (2003)).

Most AOGCMs organize too much sea-level-pressure variability into the NAM and NAO (Miller et al., 2006). The year-to-year variance of the NAM or NAO is correctly simulated by some AOGCMs, while others are significantly too variable (Osborn, 2004); for the models that simulate stronger variability, the persistence of anomalous states is greater than observed (AchutaRao et al., 2004). The magnitude of multi-decadal variability (relative to sub-decadal variability) is lower in AOGCM control simulations than is observed, and also cannot be reproduced in current model simulations with external forcings (Osborn, 2004; Gillett, 2005). However, Scaife et al. (2005) show that the observed multidecadal trend in the surface NAM and NAO can be reproduced in an AOGCM if observed trends in the lower stratospheric circulation are prescribed in the model.Troposphere-stratosphere coupling processes may therefore need to be included in models to fully simulate NAM variability. The response of the NAM and NAO to volcanic aerosols (Stenchikov et al., 2002), sea surface temperature variability (Hurrell et al., 2004) and sea-ice anomalies (Alexander et al., 2004) demonstrate some compatibility with observed variations, though the difficulties in determining cause and effect in the coupled system limit the conclusions that can be drawn with regards to the trustworthiness of model behaviour.

Like its Northern Hemisphere counterpart, the NAM, the SAM (see Chapters 3 and 9) has signatures in the tropospheric circulation, the stratospheric polar vortex, midlatitude storm tracks, ocean circulation, and sea ice. AOGCMs generally simulate the SAM realistically (Fyfe et al., 1999; Cai et al., 2003; Miller et al., 2006). For example, Figure 8.12 (adapted from Miller et al., 2006) compares the austral winter SAM simulated in the multi-model dataset at PCMDI to the observed SAM as represented in the NCEP Reanalysis. The main elements of the pattern, the low-pressure anomaly over Antarctica and the high-pressure anomalies equatorward of 60°S are all captured well by the AOGCMs. In all but two AOGCMs, the spatial correlation between the observed and simulated SAM is greater than 0.95. Further analysis shows that the SAM signature in surface temperature, such as the surface warm anomaly over the Antarctic Peninsula associated with a positive SAM event, is also captured by some AOGCMs (e.g., Delworth et al., 2006; Otto-Bliesner et al., 2006). This follows from the realistic simulation of the SAM-related circulation shown in Figure 8.12, because the surface temperature signatures of the SAM typically reflect advection of the climatological temperature distribution by the SAM-related circulation (Thompson and Wallace, 2000).

[INSERT FIGURE 8.12 HERE]

Although the spatial structure of the SAM is well simulated by the AOGCMs in the multi-model dataset at PCMDI, other features of the SAM such as the amplitude, the detailed zonal structure, and the temporal spectra do not always compare well with the NCEP Reanalysis SAM (Raphael and Holland, 2006; Miller et al., 2006). For example, Figure 8.12 shows that the simulated SAM variance (the square of the SAM amplitude) varies between 0.8 and 2.4 times the NCEP Reanalysis SAM variance. But such features vary considerably among different realizations of multiple-member ensembles (Raphael and Holland, 2006), and the temporal variability of the NCEP Reanalysis SAM does not compare well to station data (Marshall, 2003). Thus it is difficult to assess whether these discrepancies between the simulated SAM and the NCEP Reanalysis SAM point to shortcomings in the models or to shortcomings in the observed analysis.

Resolving these issues may require a better understanding of SAM dynamics. Although the SAM exhibits clear signatures in the ocean and stratosphere, its tropospheric structure can be simulated, for example, in atmospheric GCMs with a poorly resolved stratosphere and driven by prescribed SSTs (e.g., Limpasuvan and Hartmann, 2000; Cai et al., 2003). Even much simpler atmospheric models with one or two vertical levels produce SAM-like variability (Vallis et al., 2004). These relatively simple models capture the dynamics that underlie SAM variability — namely, interactions between the tropospheric jet stream and extratropical weather systems (Limpasuvan and Hartmann, 2000; Lorenz and Hartmann, 2001). Nevertheless, the ocean and stratosphere might still influence SAM variability in important ways. For example, AOGCM simulations suggest strong SAM-related impacts on ocean temperature, ocean heat transport, and sea-ice distribution (Watterson, 2001; Hall and Visbeck, 2002); suggesting a potential for air-sea interactions to influence SAM dynamics. Furthermore, observational and modelling studies (e.g., Baldwin et al., 2003; Thompson and Solomon, 2002; Gillett and Thompson, 2003) suggest that the stratosphere might also influence the tropospheric SAM, at least in austral spring and summer. Thus, an accurate simulation of stratosphere-troposphere and ocean-atmosphere coupling may still be necessary to accurately simulate the SAM.

8.4.2 Pacific Decadal Variability

Recent work suggests that the Pacific Decadal Oscillation (PDO, see Chapters 3 and 9) is the North Pacific expression of a near-global ENSO-like pattern of variability called the Interdecadal Pacific Oscillation or IPO (Power et al., 1999; Deser et al., 2004). The appearance of the IPO as the leading EOF of SST in AOGCMs that do not include interdecadal variability in natural or external forcing indicates that the IPO is an internally generated, natural form of variability. Note, however, that some AOGCMs exhibit an El Niño-like response to global warming (Cubasch et al., 2001) that can take decades to emerge (Cai and Whetton, 2000). Therefore some, though certainly not all, of the variability seen in the IPO and PDO indices might be anthropogenic in origin (Shiogama et al., 2005). The IPO and PDO can be partially understood as the residual of random interdecadal changes in ENSO activity (e.g., Power et al., 2005), reddened by the integrating effect of the upper ocean mixed layer (Newman et al., 2003; Power and Colman, 2006) and the excitation of low frequency off-equatorial Rossby waves (Power and Colman, 2006). Some of the interdecadal variability in the tropics also has an extratropical origin (e.g., Barnett et al., 1999; Hazeleger et al., 2001) and this might give the IPO a predictable component (Power et al., 2005).

AOGCMs do not seem to have difficulty in simulating IPO-like variability (e.g., Meehl and Hu, 2006; Yeh and Kirtman, 2004), even in AOGCMs that are too coarse to properly resolve equatorially-trapped waves important for ENSO dynamics. Some studies have provided objective measures of the realism of the modelled decadal variability. For example, Pierce et al. (2000) found that the ENSO-like decadal SST mode in the Pacific Ocean of their AOGCM had a pattern that gave a correlation of 0.56 with its observed counterpart. This compared with a correlation coefficient of 0.79 between the modelled and observed interannual ENSO mode. The reduced agreement on decadal time-scales was attributed to lower than observed variability in the North Pacific sub-polar gyre, over the southwest Pacific and along the western coast of North America. The latter was attributed to poor resolution of the coastal wave-guide in the AOGCM. The importance of properly resolving coastally-trapped waves in the context of simulating decadal variability in the Pacific has been raised in a number studies (e.g., Meehl and Hu, 2006). Finally, there has been little work evaluating the amplitude of Pacific decadal variability in AOGCMs. Manabe and Stouffer (1996) showed that the variability has roughly the right magnitude in their AOGCM but a more detailed investigation using recent AOGCMs with a specific focus on IPO-like variability would be useful.

8.4.3 Pacific-North American (PNA) Pattern

The PNA pattern (see Chapter 3) is commonly associated with the response to anomalous boundary forcing. However, PNA-like patterns have been simulated in atmospheric GCM experiments subjected to constant boundary conditions. Hence both external and internal processes may contribute to the formation of this pattern. Particular attention has been paid to the external influences due to SST anomalies related to ENSO episodes in the tropical Pacific, as well as those situated in the extratropical North Pacific. Internal mechanisms that might play a role in the formation of the PNA pattern include interactions between the slowly-varying component of the circulation and high-frequency transient disturbances, and instability of the climatological flow pattern. The myriad of observational and modelling studies on various processes contributing to the PNA pattern have been reviewed by Trenberth et al. (1998).

The ability of GCMs to replicate various aspects of the PNA pattern has been tested in coordinated experiments. Until several years ago, such experiments have been conducted by prescribing observed SST anomalies as lower boundary conditions for atmospheric GCMs. Particularly noteworthy are the ensembles of model runs performed under the auspices of the European PROVOST and the U.S. DSP projects. The skill of seasonal hindcasts of the participating models' atmospheric anomalies in different regions of the globe (including the PNA sector) has been summarized in a series of articles edited by Palmer and Shukla (2000). These results demonstrate that the prescribed SST forcing exerts a notable impact on the model atmospheres. The hindcast skill for the wintertime extratropical Northern Hemisphere is particularly high during the largest El Niño and La Niña episodes. However, these experiments indicate considerable variability of the responses in individual models, and among ensemble members of a given model. This large scatter of model responses suggests that atmospheric changes in the extratropics are only weakly constrained by tropical SST forcing.

The performance of the dynamical seasonal forecast system at the U.S. NCEP in predicting the atmospheric anomalies given prescribed anomalous SST forcing (in the PNA sector) has been assessed by Kanamitsu et al. (2002). During the large El Niño event of 1997–1998, the forecasts based on this system with one-month lead time are in good agreement with the observed changes in the PNA sector, with anomaly correlation scores of 0.8–0.9 (for 200 mb height), 0.6–0.8 (surface temperature) and 0.4–0.5 (precipitation). More recently, hindcast experiments have been launched using AOGCMs. The European effort was supported by the DEMETER (Development of a European Multimodel Ensemble System for Seasonal to Interannual Prediction) programme (Palmer et al., 2004). For the boreal winter season, and with hindcasts initiated in November, the model-generated PNA indices exhibit statistically significant temporal correlations with the corresponding observations. The fidelity of the PNA simulations is evident in both the multimodel ensemble means, as well as in the output from individual member models. However, the strength of the ensemble-mean signal remains low when compared with the statistical spread due to sampling fluctuations among different models, and among different realizations of a given model. The model skill is notably lower for other seasons, and longer lead times. EOF analyses of the geopotential height data produced by individual member models confirm that the PNA pattern is a leading spatial mode of atmospheric variability in these models.

Multi-century integrations have also been conducted at various institutions using the current generation of AOGCMs. Unlike the hindcasting or forecasting experiments mentioned above, these climate simulations are not aimed at reproducing specific ENSO events in the observed system. Diagnosis of the output from one such AOGCM integration indicates that the modeled ENSO events are linked to a PNA-like pattern in the upper troposphere (Wittenberg et al., 2006). The centers of action of the simulated patterns are systematically displaced 20–30 degrees of longitude west of the observed positions. This discrepancy is evidently linked to a corresponding spatial shift in the ENSO-related SST and precipitation anomaly centers simulated in the tropical Pacific. This finding illustrates that the spatial configuration of the PNA pattern in AOGCMs is crucially dependent on the accuracy of ENSO simulations in the tropics.

8.4.4 Cold Ocean-Warm Land (COWL) Pattern

The COWL pattern indicates that the oceans are relatively cold and the continents are relatively warm poleward of 40°N when the Northern Hemisphere is relatively warm. The COWL pattern results from the contrast in thermal inertia between the continents and oceans, which allows continental temperature anomalies to have greater amplitude, and thus more strongly influence hemispheric mean temperature. The COWL pattern has been simulated in climate models of varying degrees of complexity (e.g., Broccoli et al., 1998), and similar patterns have been obtained from cluster analysis (Wu and Straus, 2004a) and EOF analysis (Wu and Straus, 2004b) of Reanalysis data. In a number of studies, cold season trends in Northern

Hemisphere temperature and sea level pressure during the late 20th century have been associated with secular trends in indices of the COWL pattern (Wallace et al., 1996; Lu et al., 2004).

In their analysis of AOGCM simulations, Broccoli et al. (1998) found that the original method for extracting the COWL pattern could yield potentially misleading results when applied to a simulation forced by past and future variations in anthropogenic forcing (as is the case with most other patterns, or modes, of climate variability). The resulting spatial pattern was a mixture of the patterns associated with unforced climate variability and the anthropogenic fingerprint. Broccoli et al. (1998) also noted that temperature anomalies in the two continental centers of the COWL pattern are virtually uncorrelated, suggesting that different atmospheric teleconnections are involved in producing this pattern. Quadrelli and Wallace (2004) have recently shown that the COWL pattern can be reconstructed as a linear combination of the first two EOFs of monthly mean December–March sea level pressure. These two EOFs are the NAM and a mode closely resembling the PNA Pattern. A linear combination of these two fundamental patterns can also account for a substantial fraction of the wintertime trend in Northern Hemisphere sea level pressure during the late 20th century.

8.4.5 Atmospheric Regimes and Blocking

Weather, or climate, regimes are important factors in determining climate at various locations around the world and they can have a large impact on day-to-day variability (e.g., Plaut and Simonnet, 2001; Trigo et al., 2004; Yiou and Nogaj, 2004). GCMs have been found to simulate hemispheric climate regimes quite similar to those found in observations (Robertson, 2001; Achatz and Opsteegh, 2003; Selten and Branstator, 2004). Simulated regional climate regimes over the North Atlantic of strong similarity to the observed regimes are reported in Cassou et al. (2004), while the North Pacific regimes simulated in Farrara et al. (2000) are broadly consistent with those in observations. Since the TAR, agreement between different studies has improved regarding the number and structure of both hemispheric and sectoral atmospheric regimes, although this remains a subject of research (e.g., Wu and Straus, 2004a) and the statistical significance of the regimes has been discussed and remains an unresolved issue (e.g., Hannachi and O'Neill, 2001; Hsu and Zwiers, 2001; Stephenson et al., 2004; Molteni et al., 2006).

An important class of sectoral weather regimes are blocking events (see Chapter 3), associated with local reversals of the midlatitude westerlies. The most recent systematic intercomparison of atmospheric GCM simulations of Northern Hemisphere blocking (D'Andrea et al., 1998) was reported in the TAR. Consistent with the conclusions of this earlier study, recent studies have found that GCMs tend to simulate the location of Northern Hemisphere blocking more accurately than frequency or duration: simulated events are generally shorter and rarer than observed events (e.g., Pelly and Hoskins, 2003b). In an analysis of one of the AOGCM from the multi-model dataset at the PCMDI it was found that increased horizontal resolution combined with better physical parameterizations has led to improvements in Northern Hemisphere blocking and synoptic weather regimes over Europe. Finally, both GCM simulations and analyses of long datasets suggest the existence of considerable interannual to interdecadal variability in blocking frequency (e.g., Stein, 2000; Pelly and Hoskins, 2003a), highlighting the need for caution when assessing blocking climatologies derived from short records (either observed or simulated). Blocking events also occur in the Southern Hemisphere middle latitudes (Sinclair, 1996); no systematic intercomparison of observed and simulated Southern Hemisphere blocking climatologies has been carried out. There is also evidence of connections between North and South Pacific blocking and ENSO variability (e.g., Renwick, 1998; Chen and Yoon, 2002), and between North Atlantic blocks and sudden stratospheric warmings (e.g., Kodera and Chiba, 1995; Monahan et al., 2003); these connections have not been systematically explored in AOGCMs.

8.4.6 Atlantic Multidecadal Variability

The Atlantic Ocean exhibits considerable multidecadal variability with a timescales of about 50 to 100 years (see Chapter 3). This multidecadal variability appears to be a robust feature of the surface climate in the Atlantic region, as shown by tree ring reconstructions for the last few centuries (e.g., Mann et al., 1998). Atlantic multidecadal variability has a unique spatial pattern in the SST anomaly field, with opposite changes in the North and South Atlantic (e.g., Mestas-Nunez and Enfield, 1999; Latif et al., 2004), and this dipole pattern has been shown to be significantly correlated with decadal changes in Sahelian rainfall (Folland et al., 1986). Decadal variations in hurricane activity have also been linked to the multidecadal SST variability in the Atlantic (Goldenberg et al., 2001). AOGCMs simulate Atlantic multidecadal variability (e.g., Delworth et al., 1993; Latif, 1998 and references therein; Knight et al., 2005), and the simulated space-time structure is consistent with that observed (Delworth and Mann, 2000). The multidecadal variability simulated by the AOGCMs originates from variations of the MOC (see Section 8.3). The mechanisms, however, that control the variations of the MOC are fairly different across the ensemble of AOGCMs. In most AOGCMs, the variability can be understood as a damped oceanic eigenmode that is stochastically excited by the atmosphere. In a few other AOGCMs, however, coupled interactions between the ocean and the atmosphere appear to be more important. The relative roles of high and low latitude processes differ also from model to model. The variations of the Atlantic SST associated with the multidecadal variability appear to be predictable a few decades ahead, which has been shown by potential (diagnostic) and classical (prognostic) predictability studies. Atmospheric quantities do not exhibit predictability at decadal timescales in these studies, which supports the picture of stochastically forced variability.

8.4.7 El Niño-Southern Oscillation (ENSO)

During the last decade there has been steady progress in simulating and predicting ENSO (see Chapters 3 and 9) and the related global variability using AOGCMs (Latif et al. 2001; Davey et al., 2002; AchutaRao and Sperber, 2002). Over the last several years the parameterized physics has become more comprehensive (Gregory et al., 2000; Collins et al., 2001; Kiehl and Gent, 2004), the horizontal and vertical resolution, particularly in the atmospheric component models, has markedly increased (Guilyardi et al., 2004) and the application of observations in initializing forecasts has become more sophisticated (Alves et al., 2004). These improvements in model formulation have led to a better representation of the spatial pattern of the SST anomalies in the eastern Pacific (AchutaRao and Sperber, 2006). In fact, as an indication of recent model improvements some IPCC class models are being used for ENSO prediction (Wittenberg et al., 2006). Despite this progress, serious systematic errors in both the simulated mean climate and the natural variability persist. For example, the so-called "double Intertropical Convergence Zone (ITCZ)" problem noted by Mechoso et al. (1995; see Section 8.3.1) remains a major source of error in simulating the annual cycle in the tropics in most AOGCMs, which ultimately impacts the fidelity of the simulated ENSO. Along the equator in the Pacific the models fail to adequately capture the zonal SST gradient, the equatorial cold tongue structure is equatorially confined and extends too far too to the west (Cai et al., 2003), and typically have thermoclines that are far too diffuse (Davey et al., 2002). Most AOGCMs fail to capture the meridional extent of the anomalies in the eastern Pacific and tend to produce anomalies that extend too far into the western tropical Pacific. Most, but not all, AOGCMs produce ENSO variability that occurs on time scales considerably faster than observed (AchutaRao and Sperber, 2002), although there has been some notable progress in this regard over the last decade (AchutaRao and Sperber, 2006) in that more models are consistent with the observed time scale for ENSO (see Figure 8.13). The models also have difficulty capturing the correct phase locking between the annual cycle and ENSO. Further, some AOGCMs fail to represent the spatial and temporal structure of the El Niño-La Niño asymmetry (Monahan and Dai, 2004). Other weaknesses in the simulated amplitude and structure of ENSO variability have been discussed in Davey et al. (2002) and van Oldenborgh et al. (2005).

[INSERT FIGURE 8.13 HERE]

Current research points to some promise in addressing some of the above problems. For example, increasing the atmospheric resolution in both the horizontal (Guilyardi et al., 2004) and vertical (National Centers for Environmental Prediction Coupled Forecast System) may improve the simulated spectral characteristics of the variability, ocean parameterized physics has also been shown to significantly influence the coupled variability (Meehl et al., 2001), and continued methodical numerical experimentation into the sources of model error (e.g., Schneider, 2001) will ultimately suggest model improvement strategies.

In terms of ENSO prediction, the two biggest recent advances are: (i) the recognition that forecasts must include quantitative information regarding uncertainty (i.e., probabilistic prediction) and that verification must include skill measures for probability forecasts (Kirtman, 2003); and (ii) that a multi-model ensemble strategy may be the best current approach for adequately dealing with forecast uncertainty, e.g., Palmer et al. (2004), in which Figure 2 demonstrates that a multi-model ensemble forecast has better skill than a comparable ensemble based on a single model. Improvements in the use of data, particularly in the ocean, for initializing forecasts continues to yield enhancements in forecast skill (Alves et al., 2004); moreover, other research indicates that forecast initialization strategies that are implemented within the framework of the coupled system as opposed to the individual component models may also lead to substantial improvements in skill (Chen et al., 1995). However, basic questions regarding the predictability of SST in the tropical Pacific remain open challenges in the forecast community. For instance, it is unclear how westerly wind bursts, intra-seasonal variability or atmospheric weather noise in general, limits the predictability of ENSO (e.g., Thompson and Battisti, 2001; Kleeman et al., 2003; Flugel et al., 2004; Kirtman et al., 2005). There are also apparent decadal variations in ENSO forecast skill (Balmaseda et al., 1995; Ji et al., 1996; Kirtman and Schopf, 1998), and the sources of these variations are the subject of some debate. Finally, it remains unclear how changes in the mean climate will ultimately impact ENSO predictability (Collins et al., 2002).

8.4.8 Madden-Julian Oscillation (MJO)

The Madden-Julian Oscillation (MJO; Madden and Julian 1971) refers to the dominant mode of intraseasonal variability in the tropical troposphere. It is characterized by large-scale regions of enhanced and suppressed convection, coupled to a deep-baroclinic, primarily zonal circulation anomaly. Together, they propagate slowly eastward along the equator from the western Indian Ocean to the central Pacific and exhibit local periodicity in a broad 30–90 day range. Simulation of the MJO in contemporary coupled and uncoupled climate models remains unsatisfactory (e.g., Lin et al., 2006; Zhang, 2005). In part, we are now demanding more of the model simulations, as our understanding of the role of the MJO in the coupled atmosphere-ocean climate system expands. For instance, simulations of the MJO in models at the time of the TAR were judged using gross metrics (e.g., Slingo et al., 1996). The spatial phasing of the associated surface fluxes, for instance, are now recognized as critical for the development of the MJO and its interaction with the underlying ocean (e.g., Hendon, 2005; Zhang, 2005). Thus, while a model may simulate some gross characteristics of the MJO, the simulation may be deemed unsuccessful when the detailed structure of the surface fluxes is examined (e.g., Hendon, 2000).

Variability with MJO-characteristics (e.g., convection and wind anomalies of the correct spatial scale that propagate coherently eastward with realistic eastward phase speeds) is simulated in many contemporary models (e.g., Sperber et al., 2005; Zhang, 2005), but this variability is typically not simulated to occur often enough or with sufficient strength so that the MJO stands out realistically above the broad-band background variability (Lin et al., 2006). This under-estimation of the strength and coherence of convection and wind variability at MJO time and space scales means that many of the important climatic effects of the MJO (e.g., its impact on rainfall variability in the monsoons or the modulation of tropical cyclone development) are still poorly simulated in contemporary climate models. Simulation of the spatial structure of the MJO as it evolves through its life cycle is also problematic, with tendencies for the convective anomaly to split into double ITCZs in the Pacific and for erroneously strong convective signals to sometimes develop in the eastern Pacific ITCZ (e.g., Inness and Slingo, 2003). It has also been suggested that inadequate representation in climate models of cloud radiative interactions and/or of convection-moisture interactions may explain some of the difficulties in simulating the MJO (e.g., Lee et al., 2001; Bony and Emanuel, 2005).

Even though the MJO is probably not fundamentally a coupled ocean-atmosphere mode (e.g., Waliser et al., 1999), air-sea coupling does appear to promote more coherent eastward, and, in northern summer, northward propagation at MJO time and space scales. The interaction with an active ocean is important especially in the suppressed convective phase when SSTs are warming and the atmospheric boundary layer is recovering (e.g., Hendon, 2005). Thus, the most realistic simulation of the MJO is anticipated to be with AOGCMs. But, coupling, in general, has not been a panacea. While coupling in some models improves some aspects of the MJO, especially eastward propagation and coherence of convective anomalies across the Indian and western Pacific Oceans (e.g., Kemball-Cook et al., 2002; Inness and Slingo, 2003), problems with the horizontal structure and seasonality remain. Typically, models that show the most beneficial impact of coupling on the propagation characteristics of the MJO are also the models that possess the most unrealistic seasonal variation of MJO activity (e.g., Zhang, 2005). Unrealistic simulation of the seasonal variation of MJO activity implies that the simulated MJO will improperly interact with climate phenomena that are tied to the seasonal cycle (e.g., the monsoons and ENSO).

Simulation of the MJO is also adversely affected by biases in the mean state (see Section 8.4.7). These biases include the tendency for coupled models to exaggerate the double ITCZ in the Indian and western Pacific

Oceans, under predict the eastward extent of surface monsoonal westerlies into the western Pacific, and over predict the westward extension of the Pacific cold tongue. Together, these flaws limit development, maintenance and the eastward extent of convection associated with the MJO, thereby reducing its overall strength and coherence (e.g., Inness et al., 2003). To date, simulation of the MJO has proven to be most sensitive to the convective parameterization employed in climate models (e.g., Wang and Schlesinger, 1999; Maloney and Hartmann, 2001; Slingo et al., 2005). A consensus, though with exception (e.g., Liu et al., 2005), appears to be emerging that convective schemes based on local vertical stability and that include some triggering threshold produce more realistic MJO variability than those that convect too readily. However, some sophisticated models, with arguably the most physically based convective parameterizations, are unable to simulate reasonable MJO activity (e.g., Slingo et al., 2005).

8.4.9 Quasi-Biennial Oscillation (QBO)

The QBO (see Chapter 3) is a quasi-periodic wave-driven zonal-mean wind reversal that dominates the low-frequency variability of the lower equatorial stratosphere (3–100 hPa) and affects a variety of extratropical phenomena including the strength and stability of the wintertime polar vortex (e.g., Baldwin et al., 2001). Theory and observations indicate that a broad spectrum of vertically propagating waves in the equatorial atmosphere must be considered to explain the QBO. Realistic simulation of the QBO in GCMs therefore depends on three important conditions: (i) sufficient vertical resolution in the stratosphere to allow the representation of equatorial waves at the horizontally resolved scales of a GCM, (ii) a realistic excitation of resolved equatorial waves by simulated tropical weather, and (iii) parameterization of the effects of unresolved gravity waves. Due to the computational cost associated with the requirement of a well resolved stratosphere, the models employed for the current assessment do not generally include the QBO.

The inability of resolved wave driving to induce a spontaneous QBO in GCMs has been a long standing issue (Boville and Randel, 1992). Only recently (Takahashi, 1996, 1999; Horinouchi and Yoden, 1998; Hamilton et al., 2001) have two necessary conditions been identified that allow resolved waves to induce a QBO: high vertical resolution in the lower stratosphere (roughly 0.5 km), and a parameterization of deep cumulus convection with sufficiently large temporal variability. However, recent analysis of satellite and radar observations of deep tropical convection (Horinouchi, 2002) indicates that the forcing of a QBO by resolved waves alone requires a parameterization of deep convection with an unrealistically large amount of temporal variability. Consequently, it is currently thought that a combination of resolved and parameterized waves is required to properly model the QBO. The utility of parameterized non-orographic gravity-wave drag to force a QBO has now been demonstrated by a number of studies (Scaife et al., 2000; Giorgetta et al., 2002, 2006). Often an enhancement of input momentum flux in the tropics relative to that needed in the extratropics is required. Such an enhancement, however, depends implicitly on the amount of resolved waves and in turn the spatial and temporal properties of parameterized deep convection employed in each model (Horinouchi et al., 2003; Scinocca and McFarlane, 2004).

8.4.10 Monsoon Variability

Monsoon variability (see Chapters 3, 9 and 11) occurs over a range of temporal scales from intraseasonal to inter-decadal. Since the TAR, the ability of AOGCMs to simulate monsoon variability on intra-seasonal as well as inter-annual time scales has been examined. Lambert and Boer (2001) compared the AOGCMs that participated in the Coupled Model Intercomparison Project (CMIP), finding large errors in the simulated precipitation in the equatorial regions and in the Asian monsoon region. Lin et al. (2006) evaluated the intra-seasonal variation of precipitation in the multi-model dataset at PCMDI. They found that the intra-seasonal variance of precipitation in most AOGCMs was smaller than observed. The space-time spectra of most models have much less power than is observed, especially at periods shorter than 6 days. The speed of the equatorial waves is too fast, and the persistence of the precipitation is too long, in most of the AOGCMs. Annamalai et al (2006) examined the fidelity of the simulation of precipitation in the Asian monsoon region in the multi-model dataset at PCMDI. They found that just 6 of the 18 AOGCMs considered had a realistic simulation of climatological monsoon precipitation for the 20th century. For the former set of models the spatial correlation of the patterns of monsoon precipitation between the models exceeded 0.6, and the seasonal cycle of monsoon rainfall was simulated well. Among these models only 4 exhibited a robust ENSO-monsoon contemporaneous teleconnection. Cook and Vizy (2006) evaluated the simulation of the 20th century climate in North Africa in the multi-model dataset at PCMDI. They found that the simulation of

North African summer precipitation was less realistic than the simulation of summer precipitation over North America or Europe. In short, most AOGCMs do not simulate the spatial or intraseasonal variation of monsoon precipitation accurately. See Chapter 11 for a more detailed regional evaluation of simulated monsoon variability

8.4.11 Shorter-term Predictions Using Climate Models

Here we focus on the few results of initial value predictions made using models that are identical, or very close to, the models used in other chapters of this report for understanding and predicting climate change.

Weather prediction
Since the TAR it has been shown that climate models can be integrated as weather prediction models if they are initialized appropriately (Phillips et al., 2004). This advance appears to be due to: (i) improvements in the forecast model analyses and (ii) increases in the climate model spatial resolution. An advantage of testing a model's ability to predict weather is that some of the sub-grid scale physical processes that are parameterized in models (e.g., cloud formation, convection) can be evaluated on time-scales characteristic of those processes, without the complication of feedbacks from these processes altering the underlying state of the atmosphere (Pope and Stratton, 2002; Boyle et al., 2005; Williamson et al., 2005, Martin et al. 2006). Full use can be made of the plentiful meteorological datasets and observations from specialized field experiments. According to these studies, some of the biases found in climate simulations are also evident in the analysis of their weather forecasts. This suggests that ongoing improvements in model formulation driven primarily by the needs of weather forecasting may lead also to more reliable climate predictions.

Seasonal prediction
Verification of seasonal-range predictions provides a direct test of a model’s ability to represent the physical and dynamical processes controlling (unforced) fluctuations in the climate system. Satisfactory prediction of variations in key climate signals such as ENSO and its global teleconnections provides evidence that such features are realistically represented in long-term forced climate simulations.

A version of the HadCM3 AOGCM (known as GloSea) has been assessed for skill in predicting observed seasonal climate variations (Davey et al., 2002; Graham et al., 2005). Graham et al. (2005) analysed 43 years of retrospective 6-month forecasts (‘hindcasts’) with GloSea, run from observed ocean-land-atmosphere initial conditions. A 9-member ensemble was used to sample uncertainty in the initial conditions. Conclusions relevant to HadCM3 include: (i) the model is able to reproduce observed large-scale lagged responses to ENSO events in the tropical Atlantic and Indian Ocean SSTs; (ii) the model can realistically predict anomaly patterns in North Atlantic SSTs, shown to have important links with the North Atlantic Oscillation (NAO) and seasonal temperature anomalies over Europe.

The GFDL-CM 2.0 AOGCM has also been assessed for seasonal prediction. Twelve-month retrospective and contemporaneous forecasts were produced using a 6-member ensemble, over 15 years starting in1991. The forecasts were initialized using global ocean data assimilation (Derber and Rosati, 1989; Rosati et al., 1997) and observed atmospheric forcing, combined with atmospheric initial conditions derived from the atmospheric component of the model forced with observed SSTs. Results indicated considerable model skill out to 12 months for ENSO prediction (see http://www.gfdl.noaa.gov for summary skill scores). Global teleconnections, as diagnosed from the NCEP reanalysis (The GFDL Global Atmosphere Development Team, 2004), were evident throughout the 12 month forecasts.

8.5 Model Simulations of Extremes

Society’s perception of climate variability and climate change is, to a large extent, formed by the frequency and the severity of extremes. This is especially true if the extreme events have large and negative impacts on lives and property. As climate models’ resolution and the treatment of physical processes have improved, the simulation of extremes has also improved. Mainly because of the increased data availability (e.g., daily data, various indices, etc.), the modeling community has now examined the model simulations in greater detail and presented a comprehensive description of extreme events in the coupled models used for climate change projections.

Some extreme events, by their very nature of being smaller in scale and shorter in duration, are manifestations of either a rapid amplification, or an equilibration at a higher amplitude, of naturally occurring local instabilities. Large scale and long duration extreme events are generally due to persistence of weather patterns associated with air-sea and air-land interactions. A reasonable hypothesis might be that the coarse resolution AOGCMs might not be able to simulate local short duration extreme events. But that is not the case. Our assessment of the recent scientific literature shows perhaps surprisingly that the global statistics of the extreme events in the current climate, especially temperature, are generally well simulated by the current models (See Section 8.5.1). These models have been more successful in simulating temperature extremes than precipitation extremes.

The assessment of extremes, especially for temperature, has been done. by examining the amplitude, frequency and persistence of the following quantities: daily maximum and minimum temperature (hot days, cold days, frost days etc.), daily precipitation intensity and frequency, seasonal mean temperature and precipitation, and frequency and tracks of tropical cyclones. For precipitation, the assessment has been done either in terms of return values or extremely high rates of precipitation.

8.5.1 Extreme Temperature

Kiktev et al. (2003) compared station observations of extreme events with the simulations of an atmosphere-only GCM (HadAM3) forced by prescribed oceanic forcing and anthropogenic radiative forcing during 1950–1995. The indices of extreme events they used were those proposed by Frich et al. (2002). They found that inclusion of anthropogenic radiative forcing was required to reproduce observed changes in temperature extremes, particularly on large spatial scales. The decrease in the number of frost days in Southern Australia simulated by HadAM3 with anthropogenic forcing is in good agreement with the observations. The increase in the number of warm nights over Eurasia is poorly simulated when anthropogenic forcing is not included, but the inclusion of anthropogenic forcing improves the modelled trend patterns over western Russia and reproduces the general increase in the occurrence of warm nights over much of the Northern Hemisphere.

Meehl et al. (2004) compared the number of frost days simulated by the PCM model. The twentieth century simulations include the variations in solar, volcano, sulfate aerosol, ozone, and greenhouse gas forcing. Both model simulations and observations show that the number of frost days decreased by 2 days per decade in the western USA during the 20th century. The model simulations do not agree with observations in the southeastern USA. The model shows a decrease in the number of frost days in this region in the 20th century, while observations indicate an increase in this region. Meehl et al. (2004) argue that this discrepancy could be on account of the model's inability to simulate impact of El Niño events on the number of frost days in the southeastern USA. Meehl and Tebaldi (2004) compared the heat waves simulated by the PCM with observations. They defined a heat wave as the three consecutive warmest nights during the year. During the period 1961–1990, there is good agreement between the model and observations (NCEP reanalysis).

Kharin et al. (2005) examined the simulations of temperature and precipitation extremes for AMIP-2 models, some of which are atmospheric components of coupled models used in this assessment. They found that models simulate the temperature extremes, especially the warm extremes, reasonably well. Models have serious deficiencies in simulating precipitation extremes, particularly in the tropics.
Vavrus et al. (2006) used daily values of 20th century integrations from seven models. They defined a cold air outbreak "as an occurrence of two or more consecutive days during which the local mean daily surface air temperature is at least two standard deviations below the local wintertime mean temperature." They found that the climate models reproduce the location and magnitude of cold air outbreaks in the current climate.

Researchers have also established relationships between large scale circulation features and cold air outbreaks or heat waves. For example, Vavrus et al. (2006) found that "the favored regions of cold air outbreaks are located near and downstream from preferred locations of atmosphere blocking." Likewise, Meehl and Tebaldi (2004) found that heat waves over Europe and North America were associated with changes in the 500hPa circulation pattern.

8.5.2 Extreme Precipitation

Sun et al. (2006) investigated the intensity of daily precipitation simulated by 18 AOGCMs, including several used in this report. They found that most of the models produce light precipitation (<10 mm day^{-1}) more often than observed, too few events of heavy precipitation, and too little precipitation in heavy events (>10 mm day^{-1}). The errors tend to cancel, so that the seasonal-mean precipitation is fairly realistic (see Section 8.3).

Since the TAR, many simulations have been made with high-resolution GCMs. Iorio et al. (2004) examined the impact of model resolution on the simulation of precipitation in United States using the CCM3 GCM. They found that the high-resolution simulation produces more realistic daily precipitation statistics. The coarse resolution model had too many days with weak precipitation and not enough with intense precipitation. This tendency was partially eliminated in the high-resolution simulation, but, in the simulation at the highest resolution (T239), the high-percentile daily precipitation was still too low. This problem was eliminated when a cloud-resolving model was embedded in every grid point of the GCM.

Kimoto et al. (2005) compared the daily precipitation over Japan in an AOGCM with two different resolutions (high res. and med res. of MIROC 3.2) and found more realistic distributions with the higher resolution. Emori et al. (2005) have shown that a high-resolution AGCM (the atmospheric part of high res. MIROC 3.2) can simulate the extreme daily precipitation realistically if there is provision in the model to suppress convection when the ambient relative humidity is below 80%, suggesting that modeled extreme precipitation can be strongly parameterization dependent. Kiktev et al. (2003) compared station observations of rainfall with the simulations of the atmosphere-only GCM (HadAM3) forced by prescribed oceanic forcing and anthropogenic radiative forcing . They found that this model shows little skill in simulating changing precipitation extremes. May (2004) examined the variability and extremes of daily rainfall in the simulation of present day climate by the ECHAM4 GCM. He found that this model simulates the variability and extremes of rainfall quite well over most of India when compared to satellite-derived rainfall. The model has, however, a tendency to overestimate heavy rainfall events in central India. Durman et al. (2001) compared the extreme daily European precipitation simulated by the HadCM2 GCM with station observations. They found that the ability of the GCM to simulate daily precipitation events exceeding 15 mm per day was good but that exceeding 30 mm per day was poor. Kiktev et al. (2003) showed that the HadAM3 GCM was able to simulate the natural variability of the precipitation intensity index (annual mean precipitation divided by number of days with precipitation below 1 mm) but was not able to simulate accurately the variability in the number of wet days (the number of days in a year with precipitation above 10 mm).

Using the Palmer Drought Severity Index (PDSI), Dai at al. (2004) concluded that globally very dry or wet areas (PDSI above +3 or below –3) have increased from 20% to 38% since 1972.
In addition to simulating the short duration events like heat waves, frost days and cold air outbreaks, models have also shown success in simulating long time scale anomalies. For example, Burke et al. (2006) have shown that in the HadCM3 model, although regional distributions of wet and dry areas are not always correctly simulated, on a global basis, and at decadal timescales, the model "reproduces the observed drying trend" as defined by the Palmer Drought Severity Index if the anthropogenic forcing is included.

8.5.3 Tropical Cyclones

The spatial resolution of the coupled ocean-atmosphere models used in the IPCC assessment is generally not high enough to resolve tropical cyclones, and especially to simulate their intensity. A common approach to investigate the effects of global warming on tropical cyclones has been to utilize the SST boundary conditions from a global change scenario run to force a high resolution atmospheric GCM. That model run is then compared with a control run using the high resolution AGCM forced with specified observed SST for the current climate (Bengtsson et al., 2006; Oouchi et al., 2006; Yoshimura et al., 2006; Camargo et al., 2005; McDonald et al., 2005; Sugi et al., 2002). There are also several idealized model experiments in which a high resolution AGCM is integrated with and without a fixed global warming or cooling of SST.Another method is to embed a high resolution regional model in the lower resolution climate model (Knutson and Tuleya, 1999; Walsh et al., 2004). Projections using these methods are discussed in Chapter 10.

Bengtsson et al. (2006) have shown that the global metrics of tropical cyclones (tropical or hemispheric averages) are broadly reproduced by the ECHAM5 model, even as a function of intensity. However varying degrees of errors (in some cases substantial) in simulated tropical storm frequency and intensity have been noted in some models (e.g., The GFDL Global Atmospheric Model Development Team (GAMDT), 2004; Camargo et al., 2005; Knutson and Tuleya, 2004). The tropical cyclone simulation has been shown to be sensitive to the choice of convection parametrisation in some cases.

Oouchi et al. (2006) used one of the highest resolution (20km) atmospheric models to simulate the frequency, distribution, and intensity of tropical cyclones in the current climate. Although there were some deficiencies in simulating the geographical distribution of tropical cyclones (overprediction of tropical cyclones between 0-10S in the Indian Ocean, and underprediction between 0-10N in the western Pacific), the overall simulation of geographical distribution and frequency was remarkably good. The model could not simulate the strongest observed maximum wind speeds, and central pressures were not as low as observed, suggesting that even higher resolution may be required to simulate the most intense tropical cyclones.

8.5.4 Summary

Because most AOGCMs have coarse resolution and large-scale systematic errors, and extreme events tend to be short-lived and have smaller spatial scales, it is somewhat surprising how well the models simulate the statistics of extreme events in the current climate, including the trends during the twentieth century (see Chapter 9 for more detail). This is especially true for the temperature extremes, but intensity, frequency and distribution of extreme precipitation are less well simulated. The higher resolution models used for projections of tropical cyclone changes (Chapter 10) produce generally good simulation of the frequency and distribution of tropical cyclones, but less good simulation of their intensity. Improvements in the intensity of precipitation and tropical cyclones with increase in the resolution of AGCMs (Oouchi et al., 2006) suggests that when climate models have sufficient resolution to explicitly resolve at least the large convective systems without using parameterizations for deep convection, it is likely that simulation of precipitation and intensity of tropical cyclones will improve.

8.6 Climate Sensitivity and Feedbacks

8.6.1 Introduction

Climate sensitivity is a metric used to characterize the response of the global climate system to a given forcing. It is broadly defined as the equilibrium global mean surface temperature change following a doubling of atmospheric CO_2 concentration (see Box 10.2). Spread in model climate sensitivity is a major factor contributing to the range in projections of future climate changes (see Chapter 10) -- along with uncertainties in future emission scenarios and rates of oceanic heat uptake. As a consequence, differences in climate sensitivity between models have received close scrutiny in all four IPCC reports. Climate sensitivity is largely determined by internal feedback processes that amplify or dampen the influence of radiative forcing on climate. To assess the reliability of model estimates of climate sensitivity, one may evaluate the ability of climate models to reproduce different climate changes induced by specific forcings. These include the Last Glacial Maximum , and the evolution of climate over the last millennium and the 20th century (see Section 9.6). The compilation and comparison of climate sensitivity estimates derived from models and from observations are presented in Chapter 10 (Box 10.2). An alternative approach, which is that followed here, it to assess the reliability of key climate feedback processes known to play a critical role in the models' estimate of climate sensitivity.

Below we explain why the estimates of climate sensitivity and of climate feedbacks differ among current models (see Section 8.6.2), we summarize our understanding of the role in climate sensitivity of key radiative feedback processes associated with water vapour and lapse rate, clouds, snow and sea-ice, and we assess the treatment of these processes in the global climate models used to make projections of future climate change (Section 8.6.3). Finally we discuss how we can assess our relative confidence in the different climate sensitivity estimates derived from climate models (see Section 8.6.4). Note that climate feedbacks associated with chemical or biochemical processes are not discussed in this section (they are addressed in

Chapters 7 and 10), nor are local scale feedbacks (e.g., between soil moisture and precipitation, Section 8.2.3.2).

8.6.2 Interpretation of the Range of Climate Sensitivity Estimates Among GCMs.

8.6.2.1 Definition of Climate Sensitivity

As defined in previous assessments (Cubasch et al., 2001) and in the glossary, the global annual mean surface air temperature change experienced by the climate system after it has attained a new equilibrium in response to a CO_2 doubling is referred to as the *equilibrium climate sensitivity* (unit is K), and is often simply termed the climate sensitivity. It has long been estimated from numerical experiments in which an atmospheric GCM is coupled to a simple nondynamic model of the upper ocean with prescribed ocean heat transports (usually referred to as 'mixed-layer' or 'slab' ocean models) and the atmospheric concentration of carbon dioxide is doubled. In OAGCMs and non-steady-state (or transient) simulations, the *transient climate response* (TCR) (Cubasch et al., 2001) is defined as the global annual mean surface air temperature change (with respect to a 'control' run) averaged over a 20-year period centered at the time of CO_2 doubling in a 1%/yr compound CO_2 increase scenario. That response depends both on the sensitivity and on the ocean heat uptake. An estimate of the equilibrium climate sensitivity in transient climate change integrations is obtained from the *effective climate sensitivity* (Murphy, 1995). It corresponds to the global temperature response that would occur if the OAGCM was run to equilibrium with feedback strengths held fixed at the values diagnosed at some point of the transient climate evolution. It is computed from the oceanic heat storage, the radiative forcing and the surface temperature change (Cubasch et al., 2001; Gregory et al., 2002).

The climate sensitivity depends on the type of forcing agents applied to the climate system and on their geographical and vertical distributions (Allen and Ingram, 2002; Sausen et al., 2002; Joshi et al., 2003). As it is influenced by the nature and the magnitude of the feedbacks at work in the climate response, it also depends on the mean climate state (Boer and Yu, 2003). Some differences in climate sensitivity will also result simply from differences in the particular radiative forcing calculated by different radiation codes (see Section 10.2.1 and Section 8.6.2.3). The global annual mean surface temperature change thus presents limitations regarding the description and the understanding of the climate response to an external forcing. Indeed, the regional temperature response to a uniform forcing (and even more to a vertically or geographically distributed forcing) is highly inhomogeneous. In addition, it only considers the surface mean temperature and gives no indication of the occurrence of abrupt changes or extreme events. Despite its limitations, however, the climate sensitivity remains a useful concept because many aspects in a climate model scale well with global average temperature (although not necessarily across models), because the global mean temperature of the Earth is fairly well measured, and because it provides a simple way to quantify and compare the climate response simulated by different models to a specified perturbation. By focusing on the global scale it can also help separate the climate response from regional variability.

8.6.2.2 Why Have the Model Estimates Changed Since the TAR?

The current generation of AOGCMs covers a range of equilibrium climate sensitivity from 2.1 to 4.4°C (with a mean value of 3.2°C: see Table 8.2; Box 10.2), which is quite similar to the TAR. Yet, most climate models have undergone substantial developments since the TAR (probably more than between the SAR and the TAR), that generally involve improved parameterizations of specific processes such as clouds, boundary layer or convection (see Section 8.2). In some cases, developments have also concerned numerics, dynamical cores or the coupling to new components (ocean, carbon cycle, etc.). Developing new versions of a model so as to improve the physical basis of parameterizations or the simulation of the current climate is at the heart of modelling group activities. The rationale for these changes is generally based upon a combination of process-level tests against observations or against cloud-resolving models or large-eddy-simulation models (see Section 8.2), and on the overall quality of the model simulation (see Sections 8.3 and 8.4). These developments can, and do, affect the climate sensitivity of models.

The climate sensitivity estimates (equilibrium climate sensitivity and TCR) from the latest model version[6] used by modelling groups have increased (e.g., NCAR/CCSM, MPI/ECHAM and MRI GCMs;Hadley Centre and IPSL GCMs coupled to a slab ocean), decreased (e.g., CCSR/NIES, CSIRO and GFDL GCMs; GISS GCM coupled to a slab ocean) or remained roughly unchanged (e.g., CCCma/CGCM GCMs; GISS, Hadley Centre and IPSL AOGCMs) compared to the TAR. In some models, changes in climate sensitivity are primarily ascribed to changes in the cloud parameterization or in the representation of cloud-radiative properties (e.g., CGCM, CCSM, MRI, CCSR). However, in most models the change in climate sensitivity cannot be attributed to a specific change in the model. For instance, Johns et al. (2006) show that most of the individual changes made during the development of HadGEM1 have a small impact on the climate sensitivity, and that the global effect of the individual changes largely cancel. Also, the parameterization changes can interact non-linearly with each other so that the sum of change A and change B does not produce the same as the change A+B (e.g., Stainforth et al., 2005). Finally, the interaction among the different parameterizations of a model explains why the influence on climate sensitivity of a given change is often model dependent (see Section 8.2). For instance, the introduction of the Lock boundary layer scheme (Lock et al., 2000) to HadCM3 had a minimal impact on the climate sensitivity, in contrast to the introduction of the scheme to the GFDL model (Soden et al., 2004; Johns et al., 2006).

8.6.2.3 What Explains the Current Spread in Models' Climate Sensitivity Estimates?

As discussed in Chapter 10 and throughout the last three IPCC assessments, climate models exhibit a wide range of climate sensitivity estimates (Table 8.2). Webb et al. (2006), investigating a selection of the slab versions of models in Table 8.1, found that differences in feedbacks contribute almost three times more to the range in equilibrium climate sensitivity estimates than differences in the models' radiative forcings (the spread of models' forcing is discussed in Section 10.2).

[INSERT TABLE 8.2 HERE]

Several methods have been used to diagnose climate feedbacks in GCMs, whose strengths and weaknesses are reviewed in Stephens (2005) and Bony et al. (2006). These methods include the 'partial radiative perturbation' (PRP) approach and its variants (e.g., Colman, 2003a; Soden and Held, 2006), the use of radiative-convective models and the 'cloud radiative forcing' (CRF) method (e.g., Webb et al., 2006). Since the TAR, there has been progress in comparing the feedbacks produced by climate models in $2\times CO_2$ equilibrium experiments (Colman, 2003a; Webb et al., 2006) and in transient climate change integrations (Soden and Held, 2006). Water vapour, lapse rate, cloud and surface albedo feedback parameters, as estimated by Colman (2003a), Soden and Held (2006) and Winton (2006a), are shown on Figure 8.14.

[INSERT FIGURE 8.14 HERE]

In AOGCMs, the water vapour feedback constitutes by far the strongest feedback, with a multi-model mean and standard deviation diagnosed for the multi-model dataset at PCMDI (MMD) of 1.80 ± 0.18 Wm^{-2} K^{-1}, followed by the (negative) lapse rate feedback (-0.84 ± 0.26 Wm^{-2} K^{-1}) and the surface albedo feedback (0.26 ± 0.08 Wm^{-2} K^{-1}). The cloud feedback mean is 0.69 Wm^{-2} K^{-1} with a very large inter model spread of ± 0.38 Wm^{-2} K^{-1} (Soden and Held, 2006).

A substantial spread is apparent in the strength of water vapour feedback, that is smaller in Soden and Held (2006) than in Colman (2003a). It is not known whether this smaller spread indicates a closer consensus among current AOGCMs than among older models, differences in the methodology, or differences in the nature of climate change integrations between the two studies. In both studies, the lapse rate feedback also shows a substantial spread among models, which is explained by intermodel differences in the relative surface warming of low and high latitudes (Soden and Held, 2006). Because the water vapour and temperature responses are tightly coupled in the troposphere (see Section 8.6.3.1), models with a larger (negative) lapse-rate feedback also have a larger (positive) water vapour feedback. These act to offset each other (see Box 8.1). As a result, it is more reasonable to consider the sum of water vapour and lapse-rate feedbacks as a single quantity when analyzing the causes of intermodel variability in climate sensitivity. This

[6]Unless explicitly mentioned, GCM here refers both to AOGCM (used to estimate TCR) and AGCM coupled to a slab ocean (used to estimate equilibrium climate sensitivity).

makes intermodel differences in the combination of water vapour and lapse rate feedbacks a substantially smaller contributor to the spread in climate sensitivity estimates than differences in cloud feedback (Figure 8.14). The source of the difference in mean lapse rate feedback between the two studies is unclear, but may relate to inappropriate inclusion of stratospheric temperature response in some feedback analyses (Soden and Held, 2006).

The three studies, using different methodologies to estimate the global surface albedo feedback associated with snow and sea-ice changes, all suggest that this feedback is positive in all the models, and that its range is much smaller than that of cloud feedbacks. Winton (2006a) suggests that about three-quarters of the global surface albedo feedback arises from the Northern Hemisphere (see Section 8.6.3.3).

The diagnosis of global radiative feedbacks allows us to better understand the spread of equilibrium climate sensitivity estimates among current GCMs. In the idealised situation that the climate response to a doubling of CO_2 consisted of a uniform temperature change only, with no feedbacks operating (but allowing for the enhanced radiative cooling resulting from the temperature increase) the global warming from GCMs would be around 1.2 ^{o}C (Hansen et al., 1984; Bony et al., 2006). The water vapour feedback, operating alone on top of this would at least double the response[7]. The water vapour feedback is, however, closely related to the lapse rate feedback (see above), and the two combined result in a feedback parameter of approximately 1 $Wm^{-2}K^{-1}$, corresponding to an amplification of the basic temperature response by approximately 50%. The surface albedo feedback amplifies the basic response by about 10%, and the cloud feedback does so by 10 to 50% depending on the GCM. Note, however, that because of the inherently non-linear nature of the response to feedbacks (refer footnote 7), the final impact on sensitivity is not simply the sum of these responses. The effect of multiple positive feedbacks is that they mutually amplify each other's impact on climate sensitivity.

Using feedback parameters from Figure 8.14, it can be estimated that in the presence of water vapour, lapse rate and surface albedo feedbacks, but in the absence of cloud feedbacks, current GCMs would predict a climate sensitivity (plus/minus one standard deviation) of roughly 1.9 ± 0.15^{o}C (ignoring spread from radiative forcing differences). The mean and standard deviation of climate sensitivity estimates derived from current GCMs are larger (3.2 ± 0.7°C) essentially because the GCMs all predict a positive cloud feedback (Figure 8.14) but strongly disagree on its magnitude.

The large spread in cloud radiative feedbacks leads us to conclude that *differences in cloud response are the primary source of inter-model differences in climate sensitivity* (see discussion in Section 8.6.3.2.2). However the contributions of water vapour/lapse rate and surface albedo feedbacks to sensitivity spread are non negligible, particularly since their impact is reinforced by the mean model cloud feedback being positive and quite strong.

8.6.3 Key Physical Processes Involved in Climate Sensitivity

The traditional approach in assessing model sensitivity has been to consider water vapour, lapse rate, surface albedo and cloud feedbacks separately. Although this division can be regarded as somewhat artificial because, for example, water vapour, clouds and temperature interact strongly, it remains conceptually useful, and is consistent in approach with previous assessments. This, and the relationship between lapse rate and water-vapour feedbacks, means that we will separately address the water vapour/lapse rate feedbacks and then the cloud and surface albedo feedbacks.

8.6.3.1 Water Vapour and Lapse Rate

[7] Under these simplifying assumptions the amplification of the global warming from a feedback parameter λ (in Wm^{-2} K^{-1}) with no other feedbacks operating is $\dfrac{1}{1+\lambda/\lambda_p}$, where λ_P is the 'uniform temperature' radiative cooling response (of value approximately -3.2 Wm^{-2} K^{-1}) (Bony et al., 2006). If *n* independent feedbacks operate, λ is replaced by ($\lambda_1+\lambda_2+...\lambda_n$).

Absorption of longwave radiation increases approximately with the logarithm of water-vapour concentration, while the Clausius-Clapeyron equation dictates a near-exponential increase in moisture holding capacity with temperature. Since tropospheric and surface temperatures are closely coupled (see Section 3.4.1), these constraints predict a strongly positive water vapour feedback if RH is close to unchanged. Furthermore, the combined water vapour-lapse rate feedback is relatively insensitive to changes in lapse rate for unchanged RH (Cess, 1975) due to the compensating effects of water vapour and temperature on the OLR (refer Box 8.1). Understanding processes determining the distribution and variability in RH is therefore central to our understanding of the water vapour-lapse rate feedback. To a first approximation, GCMs indeed maintain a roughly unchanged distribution of RH under greenhouse gas (GHG) forcing. More precisely, a small but widespread RH decrease in GCMs typically reduces feedback strength slightly compared with a constant RH response (Colman, 2004; Soden and Held, 2006; Figure 8.14).

In the Planetary Boundary Layer humidity is controlled by strong coupling with the surface, and a broad-scale quasi-unchanged RH response is uncontroversial (Wentz and Schabel, 2000; Dai, 2006; Trenberth et al., 2005). Confidence in GCMs' water vapour feedback is also relatively high in the extratropics, because large scale eddies, responsible for much of the moistening throughout the troposphere, are explicitly resolved, and keep much of the atmosphere at a substantial fraction of saturation throughout the year (Stocker et al., 2001). Humidity changes in the tropical middle and upper troposphere, however, are less well understood and have more TOA radiative impact than for other regions of the atmosphere (e.g., Held and Soden, 2000; Colman, 2001). Much of the research since the TAR, then, has focused on the RH response in the tropics with emphasis on the upper troposphere (see Bony et al., 2006 for a review), and confidence in the humidity response of this region is central to our confidence in modelled water vapour feedback.

The humidity distribution within the tropical free troposphere is determined by many factors, including the detrainment of vapour and condensed water from convective systems and the large-scale atmospheric circulation. The relatively dry regions of large-scale descent play a major role in tropical longwave cooling, and changes in their area or humidity could potentially have a significant impact on water vapour feedback strength (Pierrehumbert, 1999; Lindzen et al., 2001; Peters and Bretherton, 2005). Given the complexity of processes controlling tropical humidity, however, simple convincing physical arguments on changes under global scale warming are difficult to sustain, and a combination of modelling and observational studies are needed to assess the reliability of model water vapour feedback.

In contrast to cloud feedback, a strong positive water vapour feedback is a robust feature of GCMs (Stocker et al., 2001), being found across models with many different schemes for advection, convection and condensation of water vapour. High resolution mesoscale (Larson and Hartmann, 2003) and cloud resolving models (Tompkins and Craig, 1999) run on limited tropical domains also display humidity responses consistent with strong positive feedback, although with differences in the details of upper tropospheric RH (UTRH) trends with temperature. GCM experiments have found water vapour feedback strength to be insensitive to large changes in vertical resolution, as well as convective parametrisation and advection schemes (Ingram, 2002). These modeling studies provide evidence that the free tropospheric RH response of global coupled models under climate warming is not simply an artefact of GCMs or of coarse GCM resolution, since broadly similar changes are found in a range of models of different complexity and scope. Indirect supporting evidence for model water vapour feedback strength also come from experiments which show that suppressing humidity variation from the radiation code in an AOGCM produces unrealistically low interannual variability (Hall and Manabe, 1999).

Confidence in modelled water vapour feedback is dependent upon our understanding of the physical processes important for controlling UTRH, and our confidence in their representation in GCMs. The TAR noted a sensitivity of UTRH to the representation of cloud microphysical processes in several simple modelling studies. However, other evidence suggests that the role of microphysics is limited. The observed RH field in much of the tropics can be well simulated without microphysics, but simply by observed winds while imposing an upper limit of 100% RH on parcels (Pierrehumbert and Roca, 1998; Gettelman et al., 2000; Dessler and Sherwood, 2000), or by determining a detrainment profile from clear-sky radiative cooling (Folkins et al., 2002). Evaporation of detrained cirrus condensate also does not play a major part in moistening the tropical upper troposphere (Soden, 2004; Luo and Rossow, 2004), although cirrus might be important as a water vapour sink (Luo and Rossow, 2004). Overall, these studies increase confidence in GCM water vapour feedback, since they emphasise the importance of large scale advective processes, or radiation, in which confidence in representation by GCMs is high, compared with microphysical processes, in which confidence is much lower. A significant role for microphysics in determining the distribution of changes in water vapour under climate warming cannot however yet be ruled out.

Observations provide ample evidence of *regional scale* increases and decreases in tropical UTRH in response to changes in convection (Zhu et al., 2000; Bates and Jackson, 2001; Blankenship and Wilheit, 2001; Wang et al., 2001; Chen et al., 2002; Sohn and Schmetz, 2004; Chung et al., 2004). Such changes however provide little insight into *large-scale* thermodynamic relationships, (most important for the water vapour feedback) unless considered over entire circulation systems. Recent observational studies of the tropical mean UTRH response to temperature have found results consistent with that of near unchanged RH at a variety of timescales (see Section 3.4.2.2). These include responses from interannual variability (Bauer et al., 2002; Allan et al., 2003; McCarthy and Toumi, 2004), volcanic forcing (Forster and Collins, 2004; Soden et al., 2002) and decadal trends (Soden et al., 2005), although modest RH decreases are noted at high levels on interannual timescales (Minschwaner and Dessler, 2004; Section 3.4.2.3). Seasonal variations in observed global LW trapping are also consistent with a strong positive water vapour feedback (Inamdar and Ramanathan, 1998; Tsushima et al., 2005). Note, however, that humidity responses to variability or shorter timescale forcing must be interpreted cautiously, as they are not direct analogues to that from GHG increases, because of differences in patterns of warming and circulation changes.

8.6.3.1.1 Evaluation of water vapour/lapse rate feedback processes in models

Evaluation of the humidity distribution and its variability in GCMs, while not directly testing their climate change feedbacks, can assess their ability to represent key physical processes controlling water vapour, and therefore affects our confidence in their water vapour feedback. Limitations in coverage or accuracy of radiosonde measurements or reanalyses have long posed a problem for UTRH evaluation in models (Trenberth et al., 2001; Allan et al., 2004), and recent emphasis has been on assessments using satellite measurements, along with increasing efforts to directly simulate satellite radiances in models (so as to reduce errors in converting to model level RH) (e.g., Soden et al., 2002; Allan et al., 2003; Iacono et al., 2003; Brogniez et al., 2005; Huang et al., 2005).

Major features of the mean humidity distribution are reasonably simulated in GCMs, along with the consequent distribution of OLR (see Section 8.3.1). In the important subtropical subsidence regions, models show a range of skill in representing the mean UTRH. Some large regional biases have been found (Iacono et al., 2003; Chung et al., 2004), although good agreement with satellite data has also been noted in some models for distribution and regional variability (Allan et al., 2003; Brogniez et al., 2005). Uncertainties in satellite derived datasets further complicate such comparisons, however. Skill in the reproduction of 'bimodality' in the humidity distribution at different timescales has also been found to differ between models (Zhang et al., 2003; Pierrehumbert et al., 2005), possibly associated with mixing processes and resolution. Note, however, that given the near-logarithmic dependence of longwave radiation on humidity, errors in the control climate humidity have little *direct* effect on climate sensitivity: it is the fractional change of humidity as climate changes that matters (Held and Soden, 2000).

A number of new tests of large-scale variability of UTRH have been applied to GCMs since the TAR, and have generally found skill in model simulations. Allan et al. (2003) found an AGCM forced by observed SSTs simulated interannual changes in tropical mean 6.7μm radiance (sensitive to UTRH and temperature) in broad agreement with HIRS observations over the last two decades. Minschwaner et al. (2006) analysed the interannual response of tropical mean 250 hPa RH to the mean SST of the most convectively active region in 16 AOGCMs from MMD. The mean model response (a small decrease in RH) was statistically consistent with the 215 hPa response inferred from satellite observations, when uncertainties from observations and model spread were taken into account. AGCMs have been able to reproduce global or tropical mean variations in clear sky OLR (sensitive to water-vapour and temperature distributions) over seasonal (Tsushima et al., 2005) as well as interannual and decadal (Soden, 2000; Allan and Slingo, 2002) timescales (although aerosol or greenhouse gas uncertainties and sampling differences can affect these latter comparisons; Allan et al., 2003). In the lower troposphere, GCMs can simulate global scale interannual moisture variability well (e.g., Allan et al., 2003). At a smaller scale, a number of GCMs have also shown skill in reproducing regional changes in UTRH in response to circulation changes such as from seasonal or interannual variability (e.g., Soden, 1997; Allan et al., 2003; Brogniez et al., 2005).

A further test of the response of free tropospheric temperature and humidity to surface temperature in models is how well they can reproduce interannual correlations between surface temperature and vertical humidity profiles. Although GCMs are only partially successful in reproducing regional (Ross et al., 2002) and mean tropical (Bauer et al., 2002) correlations, the marked disagreement found in previous studies (Sun and Held, 1996; Sun et al., 2001) has been shown to be in large part an artifact of sampling techniques (Bauer et al., 2002).

There have also been efforts since the TAR to test GCMs' water vapour response against that from global scale temperature changes of recent decades. One recent approach has used a long period of satellite data (1982–2004) to infer trends in UTRH. That study found an AGCM, forced by observed SSTs, was able to capture the observed global and zonal humidity trends well (Soden et al., 2005). A second approach uses the cooling following the eruption of Mt Pinatubo. Using estimated aerosol forcing, Soden et al. (2002) found a model simulated response of HIRS 6.7 μm radiance consistent with satellite observations. They also found a model global temperature response similar to that observed, but not if the water vapour feedback was switched off (although the study neglected changes in cloud cover and potential heat uptake by the deep ocean). Using radiation calculations based on humidity observations, Forster and Collins (2004) found consistency in inferred water vapour feedback strength with an ensemble of coupled model integrations , although the latitude-height pattern of the observed humidity response did not closely match any single realization. They deduced a water vapour feedback of 0.9–2.5 $Wm^{-2} K^{-1}$, a range which covers that of models under GHG forcing (see Figure 8.14). An important caveat on these studies is that climate perturbation from Pinatubo is small, not sitting clearly above natural variability (Forster and Collins, 2004). Caution is also required when comparing with feedbacks from increased GHGs, because radiative forcing from volcanic aerosol is differently distributed and occurs over shorter timescales, which can induce different changes in circulation and bias the relative land/ocean response (although a recent AOGCM study has found similar global longwave clear sky feedbacks between the two forcings; Yokohata et al., 2005). Nevertheless, comparing observed and modelled water vapour response to Mt Pinatubo constitutes one way to test model ability to simulate humidity changes induced by an external global scale forcing.

At low latitudes, GCMs show negative *lapse rate* feedback because of their tendency towards a moist adiabatic lapse rate, producing amplified warming aloft (e.g., Larson and Hartmann, 2003). At mid to high latitudes enhanced low level warming, particularly in winter, contribute a positive feedback (e.g., Colman, 2003b), and global feedback strength is dependent upon the meridional warming gradient (Soden and Held, 2006). There has been extensive testing of GCM tropospheric temperature response against observational trends for climate change detection purposes (see Section 9.4.4). Although some recent studies have suggested consistency between modelled and observed changes (e.g., Fu et al., 2004; Santer et al., 2005), debate continues as to the level of agreement, particularly in the tropics (Section 9.4.4). Regardless, if RH remains close to unchanged, the combined lapse rate and water vapour feedback is relatively insensitive to differences in lapse rate response (Cess, 1975; Allan et al., 2002; Colman, 2003a).

In the *stratosphere*, GCM water vapour response is sensitive to the location of initial radiative forcing (Joshi et al., 2003; Stuber et al., 2005). Forcing concentrated in the lower stratosphere, such as from ozone changes, invoked a positive feedback involving increased stratospheric water vapour and tropical cold point temperatures in one study (Stuber et al., 2005). However, for more homogenous forcing, such as from CO_2, stratospheric water vapour contribution to model sensitivity appears weak (Stuber et al., 2001, 2005; Colman, 2001). There is observational evidence of possible long term increases in stratospheric water vapour (Section 3.4.2.3), although it is not yet clear whether this is a feedback process. If there is a significant global mean trend associated with feedback mechanisms however, this could imply a significant stratospheric water vapour feedback (Forster and Shine, 2002).

8.6.3.1.2 Summary on water vapour and lapse fate feedbacks

Significant progress has been made since the TAR in understanding and evaluating water vapour and lapse rate feedbacks. New tests have been applied to GCMs, and have generally found skill in the representation of large-scale free tropospheric humidity responses to seasonal and interannual variability, volcanic induced cooling and climate trends. New evidence from both observations and models has reinforced the conventional view of a roughly unchanged RH response to warming. It has also increased our confidence in the ability of GCMs to simulate important features of humidity and temperature response under a range of different climate perturbations. Taken together, the evidence strongly favours a combined water vapour-lapse rate feedback of around the strength found in global climate models.

Box 8.1: Upper Tropospheric Humidity and Water Vapour Feedback

Water vapour is the most important greenhouse gas in the atmosphere. Tropospheric water vapour concentration diminishes rapidly with height, since it is ultimately limited by saturation specific humidity, which strongly decreases as temperature decreases. Nevertheless, these relatively low upper tropospheric concentrations contribute disproportionately to the ‘natural’ greenhouse effect, both because temperature contrast with the surface increases with height, and because lower down the atmosphere is nearly opaque at wavelengths of strong water vapour absorption.

In the stratosphere, there are potentially important radiative impacts due to anthropogenic sources of water vapour, such as from methane oxidation (see Section 2.3.7). In the troposphere, the *radiative forcing* due to direct anthropogenic sources of water vapour (mainly from irrigation) is negligible (see Section 2.5.6). Rather, it is the response of tropospheric water vapour to warming itself – the water vapour *feedback* – that matters for climate change. In General Circulation Models (GCMs) water vapour provides the largest positive radiative feedback (see Section 8.6.2.3): alone it roughly doubles the warming in response to forcing (such as from greenhouse gas increases). There are also possible stratospheric water vapour feedback effects due to tropical tropopause temperature changes and/or changes in deep convection (see Section 3.4.2; Section 8.6.3.1.1).

The radiative effect of absorption by water vapour is roughly proportional to the logarithm of its concentration, so it is the *fractional* change in water vapour concentration, not the absolute change, which governs its strength as a feedback mechanism. GCM calculations suggest that water vapour remains at an approximately constant fraction of its saturated value (close to unchanged *relative humidity*) under global scale warming (see Section 8.6.3.1). Under such a response, for uniform warming the largest fractional change in water vapour, and thus the largest contribution to the feedback, occurs in the upper troposphere. In addition, GCMs find enhanced warming in the tropical upper troposphere, due to changes in the lapse rate (see Section 9.4.4). This further enhances moisture changes in this region, but also introduces a partially offsetting radiative response from the temperature increase, and the net effect of the combined water vapour/lapse rate feedback is to amplify the warming in response to forcing by around 50% (Section 8.6.2.3). The close link between these processes means that water vapour and lapse rate feedbacks are commonly considered together The strength of the combined feedback is found to be robust across GCMs, despite significant inter-model differences, for example, in the mean climatology of water vapour (see Section 8.6.2.3).

Confidence in modelled water vapour feedback is thus affected by uncertainties in the physical processes controlling upper tropospheric humidity, and our confidence in their representation in GCMs. One important question is the relative contribution of large-scale advective processes (in which confidence in GCMs' representation is high) compared with microphysical processes (in which confidence is much lower) for determining the distribution and variation in water vapour. Although advection has been shown to establish the general distribution of tropical upper tropospheric humidity in the present climate (see Section 8.6.3.1), a significant role for microphysics in humidity response to climate change cannot yet be ruled out.

Difficulties in observing water vapour in the upper troposphere have long hampered both observational and modelling studies, and significant limitations remain in coverage and reliability of observational humidity data sets (see Section 3.4.2). To reduce the impact of these problems, in recent years there has been increased emphasis on the use of satellite data (such as 6.3–6.7 µm thermal radiance measurements) for inferring variations or trends in humidity, and on direct simulation of satellite radiances in models as a basis for model evaluation (see Section 3.4.2; Section 8.6.3.1.1).

Variations of upper-tropospheric water vapour have been observed across timescales from seasonal and interannual to decadal, as well as in response to external forcing (see Section 3.4.2.2). At tropics-wide scales, they correspond to roughly-unchanged relative humidity (see Section 8.6.3.1), and GCMs are generally able to reproduce these observed variations. Both column-integrated (see Section 3.4.2.1) and upper-tropospheric (see Section 3.4.2.2) specific humidity have increased over the past two decades, also consistent with roughly-unchanged relative humidity. There remains substantial disagreement between different observational estimates of lapse rate changes over recent decades, but some of these are consistent with GCM simulations (see Section 3.4.1; Section 9.4.4).

Overall, since the TAR, confidence has increased in the conventional view that the distribution of relative humidity changes little as climate warms, particularly for the upper troposphere. Confidence has also increased in the ability of GCMs to represent upper-tropospheric humidity and its variations, both free and forced. Together, upper-tropospheric observational and modelling evidence provide strong support for a combined water vapour/lapse rate feedback of around the strength found in GCMs (see Section 8.6.3.1.2).

8.6.3.2 Clouds

By reflecting the solar radiation back to space (the albedo effect of clouds) and by trapping the infrared radiation emitted by the surface and the lower troposphere (the greenhouse effect of clouds), clouds exert two competing effects on the Earth's radiation budget. These two effects are usually referred to as the shortwave (SW) and longwave (LW) components of the cloud radiative forcing (CRF). The balance between these two components depends on many factors, including macrophysical and microphysical cloud properties. In the current climate, clouds exert a cooling effect on climate (the global mean CRF is negative). In response to global warming, the cooling effect of clouds on climate might be enhanced or weakened, thereby producing a radiative feedback on climate warming (Randall et al., 2000; NRC, 2003; Zhang, 2004; Stephens, 2005; Bony et al., 2006).

In many climate models, details in the representation of clouds can substantially affect the model estimates of cloud feedback and climate sensitivity (e.g., Senior and Mitchell, 1993; Le Treut et al., 1994; Yao and Del Genio, 2002; Zhang, 2004; Stainforth et al., 2005; Yokohata et al., 2005). Moreover, the spread of climate sensitivity estimates among current models arises primarily from inter-model differences in cloud feedbacks (Colman, 2003a; Soden and Held, 2006; Webb et al., 2006; Section 8.6.2, Figure 8.14). Therefore, cloud feedbacks remain the largest source of uncertainty in climate sensitivity estimates.

In this section, we assess the evolution since the TAR in our understanding of the physical processes involved in cloud feedbacks (see Section 8.6.3.2.1), in the interpretation of the range of cloud feedback estimates among current climate models (see Section 8.6.3.2.2), and in evaluation of the model cloud feedbacks using observations (see Section 8.6.3.2.3).

8.6.3.2.1 Understanding of the physical processes involved in cloud feedbacks

The Earth's cloudiness is associated with a large spectrum of cloud types, ranging from low-level boundary-layer clouds to deep convective clouds and anvils. Understanding cloud feedbacks requires an understanding of how a change in climate may affect the spectrum and the radiative properties of these different clouds, and an estimate of the impact of these changes on the Earth's radiation budget. Moreover, since cloudy regions are also moist regions, a change in the cloud fraction matters for both the water vapour and the cloud feedbacks (Pierrehumbert, 1995; Lindzen et al., 2001). Since the TAR, there have been some advances in the analysis of physical processes involved in cloud feedbacks, thanks to the combined analysis of observations, simple conceptual models, cloud resolving models, mesoscale models and GCMs. This is reviewed in Bony et al. (2006). Major issues are presented below.

Several climate feedback mechanisms involving convective anvil clouds have been examined. Hartmann and Larson (2002) proposed that the emission temperature of tropical anvil clouds is essentially independent of the surface temperature (FAT hypothesis), and that it will thus remain unchanged during climate change. This suggestion is consistent with cloud-resolving model simulations showing that in a warmer climate, the vertical profiles of mid and upper tropospheric cloud fraction, condensate and relative humidity all tend to be displaced upward in height together with the temperature (Tompkins and Craig, 1999). However this hypothesis has not yet been tested with observations or with CRM simulations having a fine vertical resolution in the upper troposphere. The response of the anvil cloud fraction to a change in temperature remains an object of debate. Assuming that an increase with temperature of the precipitation efficiency of convective clouds could decrease the amount of water detrained in the upper troposphere, Lindzen et al. (2001) speculated that the tropical area covered by anvil clouds could decrease with rising temperature, and that would lead to a negative climate feedback (iris hypothesis). Numerous objections have been raised on various aspects of the observational evidence provided so far (Chambers et al., 2002; Del Genio and Kovari, 2002; Fu et al., 2002; Harrison, 2002; Hartmann and Michelsen, 2002; Lin et al., 2002; Lin et al., 2004), leading to a vigorous debate with the authors of the hypothesis (Bell et al., 2002; Chou et al., 2002; Lindzen et al., 2002). Other observational studies (Del Genio and Kovari, 2002; Del Genio et al., 2005a) suggest an increase of the convective cloud cover with surface temperature.

Boundary-layer clouds have a strong impact on the net radiation budget (e.g., Harrison et al., 1990; Hartmann et al., 1992) and cover a large fraction of the global ocean (e.g., Norris, 1998a,b). Understanding how they may change in a perturbed climate is thus a vital part of the cloud feedback problem. The observed relationship between low-level cloud amount and a particular measure of lower tropospheric stability (Klein and Hartmann, 1993), which has been used in some simple climate models and in some GCMs' parameterizations of boundary-layer cloud amount (e.g., NCAR CCSM3, FGOALS), led to the suggestion that a global climate warming might be associated with an increased low-level cloud cover, which would produce a negative cloud feedback (e.g., Miller, 1997; Zhang, 2004). However, variants of the lower-tropospheric stability measure, that may predict boundary-layer cloud amount as well as the Klein and Hartmann (1993) measure, would not necessarily predict an increase in low-level clouds in a warmer climate (e.g., Williams et al., 2006). Moreover, observations indicate that in regions covered by low-level clouds, the cloud optical depth decreases and the SW CRF weakens as temperature is rising (Tselioudis and Rossow, 1994; Greenwald et al., 1995; Bony et al., 1997; Del Genio and Wolf, 2000; Bony and Dufresne, 2005), but the different factors that may explain these observations are not well established. Therefore, our understanding of the physical processes that control the response of boundary-layer clouds and their radiative properties to a change in climate remains very limited.

In middle-latitudes, the atmosphere is organized in synoptic weather systems, with prevailing thick, high-top frontal clouds in regions of synoptic ascent and low-level or no clouds in regions of synoptic descent. In the northern hemisphere, several climate models report a decrease in overall extratropical storm frequency and an increase in storm intensity in response to climate warming (e.g., Carnell and Senior, 1998; Geng and Sugi, 2003), and a poleward shift of the storm tracks (Yin, 2005). Using observations and reanalyses to investigate the impact that dynamical changes such as those found by Carnell and Senior (1998) would have on the NH radiation budget, Tselioudis and Rossow (2006) suggest that the increase in storm strength would have a larger radiative impact than the decrease in storm frequency, and that this would produce increased reflection of SW radiation and decreased emission of LW radiation. However the poleward shift of the storm tracks may decrease the amount of SW radiation reflected (Tsushima et al., 2006). In addition, several studies have used observations to investigate the dependence of midlatitude cloud radiative properties on temperature. Del Genio and Wolf (2000) show that the physical thickness of low-level continental clouds decreases with rising temperature, resulting in a decrease of the cloud water path and optical thickness as temperature rises, and Norris and Iacobellis (2005) suggest that over the northern hemisphere ocean, a uniform change in surface temperature would result in decreased cloud amount and optical thickness for a large range of dynamical conditions. The sign of the climate change radiative feedback associated with the combined effects of dynamical and temperature changes on extratropical clouds is still unknown.

The role of polar cloud feedbacks in climate sensitivity has been emphasized by Holland and Bitz (2003) and Vavrus (2004). However, these feedbacks remain poorly understood.

8.6.3.2.2 Interpretation of the range of cloud feedbacks among climate models.

In 2 × CO_2 equilibrium experiments performed by mixed-layer ocean-atmosphere models as well as in transient climate change integrations performed by fully coupled ocean-atmosphere models, models exhibit a large range of global cloud feedbacks, with roughly half of the climate models predicting a more negative CRF in response to global warming, and half predicting the opposite (Soden and Held, 2006; Webb et al., 2006). Several studies suggest that the sign of cloud feedbacks may not be necessarily that of CRF changes (Zhang et al., 1994; Colman, 2003a; Soden et al., 2004), due to the contribution of clear-sky radiation changes (i.e., of water vapour, temperature and surface albedo changes) to the change in CRF. The Partial Radiative Perturbation (PRP) method, that excludes clear-sky changes from the definition of cloud feedbacks, diagnoses a positive global net cloud feedback in virtually all the models (Colman, 2003a; Soden and Held, 2006). However, the cloud feedback estimates diagnosed from either the change in CRF or the PRP method are well correlated (i.e. their relative ranking is similar), and they exhibit a similar spread among GCMs.

By decomposing the GCM feedbacks into regional components or dynamical regimes, substantial progress has been made in the interpretation of the range of climate change cloud feedbacks. The comparison of coupled ocean-atmosphere GCMs used for the climate projections presented in Chapter 10 (Bony and Dufresne, 2005), of atmospheric or slab ocean versions of current GCMs (Webb et al., 2006; Williams et al., 2006; Wyant et al., 2006), or of slightly older models (Williams et al., 2003; Bony et al., 2004; Volodin, 2004; Stowasser et al.; 2006) show that inter-model differences in cloud feedbacks are mostly attributable to the SW cloud feedback component, and that the responses to global warming of both deep convective clouds and low-level clouds differ among GCMs. Recent analyses suggest that the response of boundary-layer clouds constitutes the largest contributor to the range of climate change cloud feedbacks among current GCMs (Bony and Dufresne, 2005; Webb et al., 2006; Wyant et al., 2006). It is due both to large discrepancies in the radiative response simulated by models in regions dominated by low-level cloud cover (Figure 8.15), and to the large areas of the globe covered by these regions. However, the response of other cloud types is also important because for each model it either reinforces or partially cancels the radiative response from low-level clouds. The spread of model cloud feedbacks is substantial at all latitudes, and tends to be larger in the tropics (Bony et al., 2006; Webb et al., 2006). Differences in the representation of mixed-phase clouds and in the degree of latitudinal shift of the storm tracks predicted by the models also contribute to inter-model differences in the CRF response to climate change, in particular in the extratropics (Tsushima et al., 2006).

[INSERT FIGURE 8.15 HERE]

8.6.3.2.3 Evaluation of cloud feedbacks produced by climate models.

The evaluation of clouds in climate models has long been based on comparisons of observed and simulated climatologies of top of atmosphere radiative fluxes and total cloud amount (see Section 8.3.1). However, a good agreement with these observed quantities may result from compensating errors. Since the TAR, and partly due to the use of an ISCCP simulator (Klein and Jakob, 1999; Webb et al., 2001), the evaluation of simulated cloud fields is increasingly done in terms of cloud types and cloud optical properties (Klein and Jakob, 1999; Webb et al., 2001; Williams et al., 2003; Lin and Zhang, 2004; Weare, 2004; Zhang et al., 2005; Wyant et al., 2006). It has thus become more powerful and it constrains the models more . In addition, a new class of observational tests has been applied to GCMs, using clustering or compositing techniques, to diagnose errors in the simulation of particular cloud regimes or in specific dynamical conditions (Tselioudis et al., 2000; Norris and Weaver, 2001; Jakob and Tselioudis, 2003; Williams et al., 2003; Bony et al., 2004; Lin and Zhang, 2004; Ringer and Allan, 2004; Bony and Dufresne, 2005; Del Genio et al., 2005b; Gordon et al., 2005; Bauer and Del Genio, 2006; Williams et al., 2006; Wyant et al., 2006). An observational test focused on the global response of clouds to seasonal variations has been proposed to evaluate model cloud feedbacks (Tsushima et al., 2005), but it has not been applied to current models yet.

These studies highlight some common biases in the simulation of clouds by current models (e.g., Zhang et al., 2005). This includes the overprediction of optically thick clouds and the underprediction of optically thin low and middle-top clouds. However uncertainties remain in the observational determination of the relative amounts of the different cloud types (Chang and Li 2005). For mid-latitudes, these biases have been interpreted as the consequence of the coarse resolution of climate GCMs and their resulting inability to simulate the right strength of ageostrophic circulations (Bauer and Del Genio, 2006) and the right amount of subgrid-scale variability (Gordon et al., 2005). Although the errors in the simulation of the different cloud types may eventually compensate and lead to a prediction of the mean CRF in agreement with observations (see Section 8.3), they cast doubts on the reliability of the model cloud feedbacks. For instance, given the non linear dependence of cloud albedo on cloud optical depth, the overestimate of the cloud optical thickness implies that a change in cloud optical depth, even of the right sign and magnitude, would produce a too small radiative signature. Similarly, the underprediction of low-level and mid-level clouds presumably affects the magnitude of the radiative response to climate warming in the widespread regions of subsidence. Modelling assumptions controlling the cloud water phase (liquid, ice or mixed) are known to be critical for the prediction of climate sensitivity. However the evaluation of these assumptions is just beginning (Doutriaux-Boucher and Quaas, 2004; Naud et al., 2006). Tsushima et al. (2006) suggest that observations of the distribution of each phase of cloud water in the current climate would provide a substantial constraint on the model cloud feedbacks at middle and high latitudes.

As an attempt to assess some components of the cloud response to a change in climate, several studies have investigated the ability of GCMs to simulate the sensitivity of clouds and CRF to interannual changes in environmental conditions. When examining atmosphere-mixed-layer ocean models, Williams et al. (2006) found for instance that by considering the CRF response to a change in large-scale vertical velocity and in lower tropospheric stability, a component of the local mean climate change cloud response can be related to the present-day variability, and thus evaluated using observations. Bony and Dufresne (2005) and Stowasser and Hamilton (2006) have examined the ability of the OAGCMs of Chapter 10 to simulate the change in tropical CRF to a change in sea surface temperature, in large-scale vertical velocity, and in lower tropospheric relative humidity. They show that the models are most different and least realistic in regions of subsidence, and to a lesser extent in regimes of deep convective activity. This emphasizes the necessity to improve the representation and the evaluation of cloud processes in climate models, and especially those of boundary-layer clouds.

8.6.3.2.4 Conclusion on cloud feedbacks

Despite some advances in our understanding of the physical processes that control the cloud response to climate change and in the evaluation of some components of cloud feedbacks in current models, we are not yet able to assess which of the model estimates of cloud feedback is the most reliable. However, progress has been made in the identification of the cloud types, the dynamical regimes and the regions of the globe responsible for the large spread of cloud feedback estimates among current models. This is likely to foster more specific observational analyses and model evaluations, that will improve future assessments of climate change cloud feedbacks.

8.6.3.3 Cryosphere Feedbacks

A number of feedbacks that significantly contribute to the global climate sensitivity are due to the cryosphere. A robust feature of the response of climate models to increases in atmospheric concentrations of GHGs is the poleward retreat of terrestrial snow and sea ice, and the polar amplification of increases in lower tropospheric temperature. At the same time, the high-latitude response to increased GHG concentrations is highly variable among climate models (e.g., Holland and Bitz, 2003) and does not show substantial convergence in the latest generation of AOGCMs (Chapman and Walsh, 2005; see also Section 11.8). The possibility of threshold behaviour also contributes to the uncertainty of how the cryosphere may evolve in future climate scenarios.

Arguably the most important simulated feedback associated with the cryosphere is an increase in absorbed solar radiation resulting from a retreat of highly reflective snow or ice cover in a warmer climate. Since the TAR, some progress has been made in quantifying the surface albedo feedback associated with the cryosphere. Hall (2004) found that the albedo feedback was responsible for about half the high-latitude response to a doubling of CO_2. However, an analysis of long control simulations showed that it accounted for surprisingly little internal variability. Hall and Qu (2006) show that biases of a number of MMD models in reproducing the observed seasonal cycle of land snow cover (especially the springtime melt) are tightly related to the large variations in snow albedo feedback strength simulated by the same models in climate change scenarios. Addressing the seasonal cycle biases would therefore provide a constraint that would reduce divergence in simulations of snow albedo feedback in climate change. However possible use of seasonal snow-albedo feedback to evaluate snow-albedo feedback under climate change conditions is of course dependent upon the realism of the correlation between the two feedbacks suggested by GCMs (Figure 8.16). A new result found independently by Winton (2006a) and Qu and Hall (2005) is that surface processes are the main source of divergence in climate simulations of surface albedo feedback, rather than simulated differences in cloud fields in cryospheric regions.

[INSERT FIGURE 8.16 HERE]

Our understanding of other feedbacks associated with the cryosphere, e.g. ice insulating feedback, MOC/SST-sea-ice feedback, ice-thickness/ice-growth feedback, has improved since the TAR (see for details NRC, 2003; Bony et al., 2006) . However, the relative influence on climate sensitivity of these feedbacks has not been quantified.

Understanding and evaluating sea-ice feedbacks is complicated by their strong coupling to processes in the high-latitude atmosphere and ocean, particularly to polar cloud processes and ocean heat and freshwater transport. Additionally, while impressive advances have occurred in developing sea-ice components of the AOGCMs since the TAR, particularly by the inclusion of more sophisticated dynamics in most of them (see Section 8.2.4), evaluation of cryospheric feedbacks through the testing of model parameterizations against observations is hampered by the scarcity of observational data in the polar regions. In particular, the lack of sea ice thickness observations is a considerable problem.

The role of sea-ice dynamics in climate sensitivity has remained uncertain for years. Some recent results with AGCMs coupled to slab ocean models (Hewitt et al., 2001; Vavrus and Harrison, 2003) support the hypothesis that a representation of sea-ice dynamics in climate models has a moderating impact on climate sensitivity. However, experiments with full AOGCMs (Holland and Bitz, 2003) show no compelling relationship between the transient climate response and the presence or absence of ice dynamics, with numerous model differences presumably overwhelming whatever signal might be due to ice dynamics. A substantial connection between the initial (i.e., control) simulation of sea-ice and the response to GHG forcing (Holland and Bitz, 2003; Flato, 2004) further hampers "clean" experiments aimed at identifying or quantifying the role of sea-ice dynamics.

A number of processes, other than surface albedo feedback, have been shown to also contribute to the polar amplification of warming in models (Alexeev, 2003; Alexeev et al., 2005; Cai, 2005; Holland and Bitz, 2003; Vavrus, 2004; Winton, 2006b). An important one is additional poleward energy transport, but contributions from local high latitude , water vapour, cloud and temperature feedbacks have also been found. The processes and their interactions are complex, however, with substantial variation between models (Winton, 2006b), and their relative importance contributing to or dampening high latitude amplification has not yet been properly resolved.

8.6.4 How to Assess Our Relative Confidence in the Feedbacks Simulated by the Different Models?

Assessments of our relative confidence in climate projections from the different models should ideally be based on a comprehensive set of observational tests that would allow us to quantify model errors in simulating a wide variety of climate statistics, including simulations of the mean climate and variability, and of particular climate processes. The collection of measures that quantify how well a model performs in an ensemble of tests of this kind are referred to as "*climate metrics*". To have the ability to constrain future climate projections, they would ideally have strong connections with one or several aspects of climate change: climate sensitivity, large-scale patterns of climate change (interhemispheric symmetry, polar amplification, vertical patterns of temperature change, land-sea contrasts), regional patterns, or transient aspects of climate change. For example, to assess our confidence in model projections of the Australian climate, one would need in the metrics some measures of the quality of ENSO simulation because the Australian climate depends much on this variability (see Section 11. 7).

To better assess our confidence in the different model estimates of climate sensitivity, two kinds of observational tests are available: tests related to the global climate response associated with specified external forcings (discussed in Chapters 6, 9 and 10; Box 10.2), and tests focused on the simulation of key feedback processes.

Based on our understanding of both the physical processes that control key climate feedbacks (see Section 8.6.3), and also the origin of intermodel differences in the simulation of feedbacks (see Section 8.6.2), the following climate characteristics appear to be particularly important: (i) for the water vapor and lapse rate feedbacks, the response of upper tropospheric relative humidity and lapse rate to interannual or decadal changes in climate; (ii) for cloud feedbacks, the response of boundary-layer clouds and anvil clouds to a change in surface or atmospheric conditions and the change in cloud radiative properties associated with a change in extratropical synoptic weather systems; (iii) for snow-albedo feedbacks, the relationship between surface air temperature and snow melt over northern land areas during springtime; and (iv) for sea-ice feedbacks, the simulation of sea-ice thickness.

A number of diagnostic tests have been proposed since the TAR (see Section 8.6.3), but few of them have been applied to a majority of the models currently in use. Moreover, it is not yet clear which tests are critical for constraining future projections. Consequently, a set of model metrics that might be used to narrow the range of plausible climate change feedbacks and climate sensitivity has yet to be developed.

8.7 Mechanisms Producing Thresholds and Abrupt Climate Change

8.7.1 Introduction

Our discussion of thresholds and abrupt climate change is based on the definitions of "threshold" and "abrupt" proposed by Alley et al. (2002). The climate system tends to respond to changes in a gradual way until it crosses some threshold: thereafter any change that is defined as abrupt is one where the change in the response is much larger than the change in the forcing. The changes at the threshold are therefore abrupt relative to the changes that occur before or after the threshold and can lead to a transition to a new state. The space scales for these changes can range from global to local. In this definition, the magnitude of the forcing and response are important. In addition to the magnitude, the time scale being considered is also important. Here we mainly focus on the decadal to centennial time scales.

Because of the somewhat subjective nature of the definitions of threshold and abrupt, there have been efforts to develop quantitative measures to identify these points in a time series of a given variable (e.g., Lanzante, 1996; Seidel and Lanzante, 2004; Tomé and Miranda, 2004). The most common way to identify thresholds and abrupt changes is by linearly detrending the input time series and looking for large deviations from the trend line. More statistically rigorous methods are usually based on Bayesian statistics.

Here we explore the potential causes and mechanisms for producing thresholds and abrupt climate change and address the issue of how well climate models can simulate these changes. The following discussion is split into two main areas: forcing changes that can result in abrupt changes and abrupt climate changes that result from large natural variability on long time scales. Formally the latter abrupt changes do not fit the definition of thresholds and abrupt changes, because the forcing (at least radiative forcing - the external boundary condition) is not changing in time. However these changes have been discussed in the literature and popular press and are worthy of assessment here.

8.7.2 Forced Abrupt Climate Change

8.7.2.1 Meridional Overturning Circulation Changes

As the radiative forcing of the planet changes, the climate system responds on many different time scales. For the physical climate system typically simulated in coupled models (atmosphere, ocean, land, sea ice), the longest response time scales are found in the ocean (Stouffer, 2004). In terms of thresholds and abrupt climate changes on decadal and longer time scales, the ocean has also been a focus of attention. In particular, the ocean's Atlantic meridional overturning circulation (MOC, see Box 5.1 for definition and description) is a main area of study.

The MOC transports large amounts of heat (order of 10^{15} watts) and salt into high latitudes of the North Atlantic. There, the heat is released to the atmosphere, cooling the surface waters. The cold, relatively salty waters sink to depth and flow southward out of the Atlantic basin. The complete set of climatic drivers of this circulation remains unclear but it is likely that both density (e.g., Stommel 1961; Rooth 1982) and wind stress forcings (e.g., Wunsch, 2002; Timmermann and Goosse, 2004) are important. Both paleo-studies (e.g., Broecker, 1997; Clark et al., 2002) and modeling studies (e.g., Manabe and Stouffer, 1988, 1997; Vellinga and Wood, 2002) suggest that disruptions in the MOC can produce abrupt climate changes. A systematic model intercomparison study (Rahmstorf et al., 2005) found that all 11 participating Earth system models of intermediate complexity have a threshold where the MOC shuts down (See Section 8.8.3). Due to the high computational cost, such a search for thresholds has not yet been performed with AOGCMs .

It is important to note the distinction between the equilibrium and transient or time-dependent responses of the MOC to changes in forcing. Due to the long response time scales found in the ocean (some longer than 1000 years), it is possible that the short term response to a given forcing change may be very different from the equilibrium response. Such behavior of the coupled system has been documented in at least one AOGCM (Stouffer and Manabe, 2003) and suggested in the results of a few other AOGCM studies (e.g., Hirst, 1999; Senior and Mitchell, 2000, Bryan et al. 2006). In these AOGCM experiments, the MOC weakens as the greenhouse gases increase in the atmosphere. When the CO_2 concentration is stabilized, the MOC slowly returns to its unperturbed value.

As discussed in section 10.3.4, the MOC typically weakens as greenhouse gases increase due to the changes in surface heat and freshwater fluxes in high latitudes (Manabe et al., 1991). The surface flux changes cause the surface density to reduce, hindering the vertical movement of water, slowing the MOC. As the MOC slows, it could approach a threshold where the circulation can no longer sustain itself. Once the MOC crosses this threshold, it could rapidly change states, causing abrupt climate change where the North Atlantic and surrounding land areas would cool relative to the case where the MOC is active. This cooling is the result of the loss of heat transport from low latitudes in the Atlantic and the feedbacks associated with the reduction in the vertical mixing of high latitude waters.

A common misunderstanding is that the MOC weakening could cause the onset of an ice age. However, no model has supported this speculation when forced with realistic estimates of future climate forcings (see Section 10.3.4). In addition, idealized modeling studies where the MOC was forced to shut down through very large sources of freshwater (not changes in GHG), the surface temperature changes do not support the idea that an ice age could result from a MOC shut down, though the impacts on climate would be large (Manabe and Stouffer, 1988, 1997; Schiller et al., 1997; Vellinga and Wood, 2002; Stouffer et al., 2006). In a recent intercomparison involving 11 coupled atmosphere-ocean models (Gregory et al., 2005), the MOC decreases by only 10–50% during a 140-year period (as CO_2 quadruples), and in no model is there a land cooling anywhere (as the global-scale heating due to increasing CO_2 overwhelms the local cooling effect due to reduced MOC).

Because of the large amount of heat and salt transported northward and its sensitivity to surface fluxes, the changes in the MOC are able to produce abrupt climate change on decadal to centennial time scales (e.g., Manabe and Stouffer, 1995; Stouffer et al., 2006). Idealized studies using present day simulations have shown that models can simulate many of the variations seen in the paleo-record on decadal to centennial time scales when forced by fluxes of freshwater water at the ocean surface. However, the quantitative response to freshwater inputs varies widely among models (Stouffer et al., 2006) which led the Coupled Model intercomparison Project (CMIP) and Paleo-Model Intercomparison Project (PMIP) panels to design and support a set of coordinated experiments to study this issue (http://www.gfdl.noaa.gov/~kd/CMIP.html and http://www-lsce.cea.fr/pmip/).

In addition to the amount of the freshwater input, the exact location may also be important (Rahmstorf 1996, Manabe and Stouffer, 1997; Rind et al., 2001). Designing experiments and determining the realistic past forcings needed to test the models' response on decadal to centennial time scales, remains to be accomplished.

The processes determining MOC response to increasing GHG have been studied in a number of models. In many models, initial MOC response to increasing GHG is dominated by thermal effects. In most models this is enhanced by changes in salinity driven by, among other things, the expected strengthening of the hydrological cycle (Gregory et al., 2005; Chapter 10). Meltwater runoff from a melting of the Greenland ice sheet is a potentially major source of freshening not yet included in the models found in the multi-model dataset (see Section 8.7.2.2). More complex feedbacks, associated with wind and hydrological changes, are also important in many models. These include local surface flux anomalies in deep water formation regions (Gent, 2001), and oceanic teleconnections driven by changes to the fresh water budget of the tropical and South Atlantic (e.g., Latif et al., 2000; Thorpe et al., 2001; Vellinga et al., 2002; Hu et al., 2004). The magnitudes of the climate factors causing the MOC to weaken, along with the feedbacks and the associated restoring factors, are all uncertain at this time. Evaluation of these processes in AOGCMs is mainly restricted by lack of observations, but some early progress has been made in individual studies (e.g., Schmittner et al., 2000; Pardaens et al., 2003; Wu et al., 2005; Chapter 9). Model intercomparison studies (e.g., Gregory et al., 2005; Stouffer et al., 2006; Rahmstorf et al. 2005) were developed to identify and understand the causes for the wide range of MOC responses in the coupled models used here (see Chapters 4, 6 and 10).

8.7.2.2 Rapid West Antarctic and/or Greenland Ice Sheet Collapse and MOC Changes

Increased influx of freshwater to the ocean from the ice sheets is a potential forcing for abrupt climate changes. For Antarctica in the present climate, these fluxes chiefly arise from melting below the ice shelves and from melting of icebergs transported by the ocean; both fluxes could increase significantly in a warmer climate. Ice sheet runoff and iceberg calving, in roughly equal shares, currently dominate the freshwater flux from the Greenland ice sheet (Church et al., 2001; Chapter 4). In a warming climate, runoff is thought to quickly increase and become much larger than the calving rate, the latter of which in turn is likely to decrease as less and thinner ice borders the ocean; basal melting from below the grounded ice will remain several orders of magnitude smaller than the other fluxes (Huybrechts et al., 2002). For a discussion of the likelihood of these ice sheet changes and the effects on sea level, see the discussion in Chapter 10.

Changes in the surface forcing near the deepwater production areas seem to be most capable of producing rapid climate changes on decadal and longer time scales due to changes in the ocean circulation and mixing. If there are large changes in the ice volume over Greenland, it is likely that much of this meltwater will freshen the surface waters in the high latitude N Atlantic, slowing down the MOC (see Section 8.7.2.1; Chapter 10). Rind et al. (2001) found that changes in the NADW formation rate could instigate changes in the deepwater formation around Antarctica.

The response of the Atlantic MOC to changes in the Antarctic ice sheet is less well understood. Experiments with ocean-only models where the meltwater changes are imposed as surface salinity changes, indicate that the Atlantic MOC will intensify as the waters around Antarctica become lighter (Seidov et al., 2001). Weaver et al. (2003) showed that by adding freshwater in the Southern Ocean, the MOC could change from an "off" state to a state similar to present day. However, in an experiment with an AOGCM, Seidov et al. (2005) found that an external source of freshwater in the Southern Ocean resulted in a surface freshening throughout the world ocean, weakening the Atlantic MOC. In these model results, the Southern Hemisphere MOC associated with Antarctic Bottom Water (AABW) formation weakened, causing a cooling around Antarctica. See Chapters 4, 6 and 10 for more discussion on the likelihood of large meltwater fluxes from the icesheets impacting the climate.

In summary, there is a potential for rapid ice sheet changes to produce rapid climate change both through sea level changes and ocean circulation changes. The ocean circulation changes result from increased freshwater flux over the particularly sensitive deep water production sites. In general, the possible climate changes associated with future evolution of the Greenland Ice Sheet are better understood than those associated with changes in the Antarctic Ice Sheets.

8.7.2.3 Volcanoes

Volcanoes produce abrupt climate responses on short time scales. The surface cooling effect of the stratospheric aerosols, the main climatic forcing factor, decays in 1 to 3 years after an eruption due to the lifetime of the aerosols in the stratosphere. It is possible for one large volcano or a series of large volcanic eruptions to produce climate responses on longer time scales, especially in the subsurface region of the ocean (Gleckler et al., 2006b; Delworth et al., 2005).

The models' ability to simulate any possible abrupt response of the climate system to volcanic eruptions seems conceptually similar to their ability to simulate the climate response to future changes in GHG in that both produce changes in the radiative forcing of the planet. However, mechanisms involved in the exchange of heat between the atmosphere and ocean may be different in response to volcanic forcing when compared to the response to increase GHG. Therefore the feedbacks involved may be different (see Section 9.6.2.2 for more discussion).

8.7.2.4 Methane Hydrate Instability/Permafrost Methane

Methane hydrates are stored on the sea bed along continental margins where they are stabilized by high pressures and low temperatures, implying that ocean warming may cause hydrate instability and release of methane into the atmosphere (see Section 4.7.2.4). Methane is also stored in the soils in areas of permafrost and warming increases the likelihood of a positive feedback in the climate system via permafrost melting and the release of trapped methane into the atmosphere. The likelihood of methane release from methane hydrates found in the oceans or methane trapped in permafrost layers is assessed in Chapter 7.

Here we consider the potential usefulness of models in determining if those releases could trigger an abrupt climate change. Both forms of methane release represent a potential threshold in the climate system. As the climate warms, the likelihood of the system crossing a threshold for a sudden release increases (see Chapters 4, 7, 10). Since these changes produce changes in the radiative forcing through changes in the GHG concentrations, the climatic impacts of such a release are the same as an increase in the rate of change in the radiative forcing. Therefore the models ability to simulate any abrupt climate change should be similar to their ability to simulate future abrupt climate changes due to changes in the GHG forcing.

8.7.2.5 Biogeochemical

Two questions concerning biogeochemical aspects of the climate system will be addressed here. One is: can biogeochemical changes lead to abrupt climate change? The second is: can abrupt changes in the MOC further impact the radiative forcing through biogeochemical feedbacks?

Abrupt changes in biogeochemical systems of relevance to our capacity to simulate the climate of the 21st Century are not well understood (Friedlingstein et al., 2003). The potential for major abrupt change exists in the uptake and storage of carbon by terrestrial systems. While abrupt change within the climate system is beginning to be seriously considered (Rial et al., 2004; Schneider, 2004) the potential for abrupt change in terrestrial systems, such as loss of soil carbon (Cox et al., 2000) or die-back of the Amazon forests (Cox et al., 2004) remains uncertain. In part this is due to lack of understanding of processes (see Friedlingstein et al., 2003; Chapter 7) and in part it results from the impact of differences in the projected climate sensitivities in the host climate models (Joos et al., 2001; Govindasamy et al., 2005; Chapter 10) where changes in the physical climate system impact the biological response.

There is some evidence of multiple equilibria within vegetation-soil-climate systems. These include North Africa and Central East Asia where Claussen (1998) using a EMIC with a land-vegetation component, showed two stable equilibria for rainfall, dependent on initial land surface conditions. Kleidon et al. (2000), Wang and Eltahir (2000) and Renssen et al. (2003) also found evidence for multiple equilibria. These are preliminary assessments using relatively simple physical climate models that highlight the possibility of irreversible change in the Earth System but require extensive further research to assess the reliability of the phenomena found.

There have only been a few preliminary studies of the impact of abrupt climate changes such as the shutdown of the MOC on the carbon cycle. The findings of these studies indicate that the shutdown of the MOC would tend to increase the amount of GHG in the atmosphere (Joos et al., 1999; Plattner et al., 2001; Chapter 6). In both these studies, only the effect of oceanic component of the carbon cycle changes was considered.

8.7.3 Unforced Abrupt Climate Change

Formally, as noted above, the changes discussed here do not fall into the definition of abrupt climate change. In the literature, unforced abrupt climate change falls into two general categories. One is just a red noise time series, where there is power at decadal and longer time scales. A second category is a bimodal or multi-modal distribution. In practice, it can be very difficult to distinguish between the two categories unless the time series are very long—long enough to eliminate sampling as an issue—and the forcings are fairly constant in time. In observations, neither of these conditions is normally met.

Models, both AOGCMs and less complex models, have produced examples of large abrupt climate change (e.g., Hall and Stouffer 2001; Goosse et al. 2002) without any changes in forcing. Typically, these events are associated with changes in the ocean circulation, mainly in the N Atlantic. An abrupt event can last for several years to a few centuries. They bear some similarities with the conditions observed during a relatively cold period in the recent past in the Arctic (Goosse et al., 2003)

Unfortunately, the probability of such an event is difficult to estimate as it requires a very long experiment and is certainly dependent on the mean state simulated by the model. Furthermore, comparison with observations is nearly impossible since it would require a very long period with constant forcing which does not exist in nature. Nevertheless, if an event such as the one of those mentioned above were to occur in the future, it would make the detection and attribution of climate changes very difficult.

8.8 Representing the Global System with Simpler Models

8.8.1 Why Lower Complexity?

An important concept in climate system modelling is the notion of a spectrum of models of differing levels of complexity, each being optimum for answering specific questions. It is not meaningful to judge one level as being better or worse than another independently of the context of analysis. What is important is that each model be asked questions appropriate for its level of compexity and quality of its simulation.

The most comprehensive models available are AOGCMs. These models, which include more and more components of the climate system (see Section 8.2), are designed to provide the best representation of the system and its dynamics, thereby serving as the most realistic laboratory of nature. Their major limitation is their high computational cost. To date, unless modest resolution models are executed on an exceptionally large-scale distributing computed system, as in the climate*prediction*.net project (Stainforth et al., 2005), only a limited number of multi-decadal experiments can be performed with AOGCMs, which hinders a systematic exploration of uncertainties in climate change projections and prevents studies of the long-term evolution of climate.

At the other end of the spectrum of complexity of climate system models are the so-called simple climate models (see Harvey et al., 1997 for a review of these models). The most advanced simple climate models contain modules that calculate in a highly parameterised way (1) the abundances of atmospheric greenhouse gases for given future emissions, (2) the radiative forcing resulting from the modelled greenhouse gas concentrations and aerosol precursor emissions, (3) the global mean surface temperature response to the computed radiative forcing and (4) the global mean sea level rise due to thermal expansion of sea water and the response of glaciers and ice sheets. These models are much more computationally efficient than AOGCMs and thus can be utilised to investigate future climate change in response to a large number of different scenarios of greenhouse gas emissions. Uncertainties from the modules can also be concatenated, potentially allowing the climate and sea level results to be expressed as probabilistic distributions, which is harder to do with AOGCMs because of their computational expense. A characteristic of simple climate models is that climate sensitivity and other subsystem properties must be specified based on the results of AOGCMs or observations. Therefore, simple climate models can be tuned to individual AOGCMs and employed as a tool to emulate and extend their results (e.g., Raper et al., 2001; Cubasch et al., 2001). They are useful mainly for examining global-scale questions.

To bridge the gap between AOGCMs and simple climate models, Earth system models of intermediate complexity (EMICs) have been developed. Given that this gap is quite large, there is a wide range of EMICs (see the reviews of Saltzman, 1978 and Claussen et al., 2002). Typically, EMICs use a simplified atmospheric component coupled to an OGCM or simplified atmospheric and oceanic components. The degree of simplification of the component models varies from EMIC to EMIC.

EMICs are reduced-resolution models that incorporate most of the processes represented by AOGCMs, albeit in a more parameterised form. They explicitly simulate the interactions between various components of the climate system. Similarly to AOGCMs, but in contrast to simple climate models, the number of degrees of freedom of an EMIC exceeds the number of adjustable parameters by several orders of magnitude. However, these models are simple enough to permit climate simulations over several thousand of years or even glacial cycles (with a period of some 100,000 years), although not all are suitable for this purpose. Moreover, like simple climate models, EMICs can explore the parameter space with some completeness and are thus appropriate for assessing uncertainty. EMICs can also be utilised to screen the phase space of climate or the history of climate in order to identify interesting time slices, thereby providing guidance for more detailed studies to be undertaken with AOGCMs. Besides, EMICs are invaluable tools for understanding large-scale processes and feedbacks acting within the climate system. Certainly, it would not be sensible to apply an EMIC to studies which require high spatial and temporal resolution. Furthermore, model assumptions and restrictions, hence the limit of applicability of individual EMICs, must be carefully studied. Some EMICs include a zonally averaged atmosphere or zonally averaged oceanic basins. In a number of EMICs, cloudiness and/or wind fields are prescribed and do not evolve with changing climate. In still other EMICs, the atmospheric synoptic variability is not resolved explicitly, but diagnosed by using a statistical-dynamical approach. A priori, it is not obvious how the reduction in resolution or dynamics/physics affects the simulated climate. As shown in Section 8.8.3 and in Chapters 6, 9 and 10, at large scale, most EMIC results compare well with observational or proxy data and AOGCM results. Therefore, it is argued that there is a clear advantage in having available a spectrum of climate system models.

8.8.2 Simple Climate Models

As in the TAR, a simple climate model is utilised in this report to emulate the projections of future climate change conducted with state-of-the-art AOGCMs, thus allowing the investigation of the temperature and sea level implications of all relevant emission scenarios (see Chapter 10). This model is an updated version of the MAGICC model (Wigley and Raper, 1992, 2001; Raper et al., 1996). The calculation of the radiative forcings from emission scenarios closely follows that described in Chapter 2, and the feedback between climate and the carbon cycle is treated consistently with Chapter 7. The atmosphere-ocean module consists of an atmospheric energy balance model coupled to an upwelling-diffusion ocean model. The atmospheric energy balance model has land and ocean boxes in each hemisphere, and the upwelling-diffusion ocean model in each hemisphere has 40 layers with inter-hemispheric heat exchange in the mixed layer.

This simple climate model has been tuned to outputs from 19 of the AOGCMs described in Table 8.1, with resulting parameter values as given in the Supplementary Material, Table S8.1. The applied tuning procedure involves an iterative optimisation to derive least-square optimal fits between the simple model results and the AOGCM outputs for temperature time series and net oceanic heat uptake. This procedure attempts to match not only the global mean temperature but also the hemispheric and land and ocean surface temperature changes of the AOGCM results by adjusting the equilibrium land-ocean warming ratio. Where data availability allowed, the tuning procedure took simultaneously account of lowpass filtered AOGCM data for two scenarios, namely a 1% per year compounded increase in CO_2 concentration to twice (four times) the pre-industrial level, with subsequent stabilization. Before tuning, the AOGCM temperature and heat uptake data has been de-drifted by substracting the respective lowpass-filtered pre-industrial control run segments. The three tuned parameters in the simple climate model are the effective climate sensitivity, ΔT_{eff}, the ocean effective vertical diffusivity, k, and the equilibrium land-ocean warming ratio, RLO. AOGCM specific values for the radiative forcing for CO_2 doubling, $F_{2\times}$, were used in the tuning procedure where available (from Forster and Taylor, 2006, supplemented with values provided directly from the modelling groups). Otherwise, a default value of 3.71 W m^{-2} was chosen (Myhre et al., 1998). Default values of 1 W m^{-2} $°C^{-1}$, 1 W m^{-2} $°C^{-1}$ and 8°C were used for the land-ocean heat exchange coefficient, LO, the inter-hemispheric heat exchange coefficient, NS, and the magnitude of the warming that would result in a collapse of the MOC, ΔT^{+} (see TAR, Chapter 9, Appendix 9.1).

The obtained best-fit climate sensitivity estimates differ for various reasons from other estimates that are derived with alternative methods. Such alternative methods are for example regression estimates that use a global energy balance equation around the year of CO_2 doubling or the analysis of slab ocean equilibrium warmings. The resulting differences in climate sensitivity estimates can be partially explained by the non-time constant effective climate sensitivities in many of the AOGCM runs. Furthermore, tuning results of a simple climate model will be affected by the model structure, although simple, and other default parameter settings that affect the simple model transient response.

8.8.3 Earth System Models of Intermediate Complexity

Pictorially, EMICs can be defined in terms of the components of a three-dimensional vector (Claussen et al., 2002): the number of interacting components of the climate system explicitly represented in the model, the number of processes explicitly simulated and the detail of description. Some basic information on the EMICs used in Chapter 10 of this report is presented in Table 8.3. A comprehensive description of all EMICs in operation can be found in Claussen (2005). Actually, there is a broad range of EMICs, reflecting the differences in scope. In some EMICs, the number of processes and the detail of description are reduced to simulate feedbacks between as many components of the climate system as feasible. Others, with less interacting components, are utilised in long-term ensemble experiments to investigate specific aspects of climate variability. The gap between some of the most complicated EMICs and AOGCMs is not so large. In fact, this particular class of EMICs is derived from AOGCMs. On the other hand, EMICs and simple climate models differ much more. For instance, EMICs as well as AOGCMs realistically represent the large-scale geographical structures of the Earth, like the shape of continents and oceanic basins, which is certainly not the case for simple climate models.

[INSERT TABLE 8.3 HERE]

Since the TAR, EMICs have intensively been used to study past and future climate changes (see Chapters 6, 9 and 10). Furthermore, a great deal of effort has been devoted to the evaluation of those models through coordinated intercomparisons.

Figure 8.17 compares results of some of the EMICs utilised in Chapter 10 (see Table 8.3) with observation-based estimates and results of GCMs that took part in AMIP (Atmospheric Model Intercomparison Project) and CMIP1 (Coupled Model Intercomparison Project, phase 1) (Gates et al., 1999; Lambert and Boer, 2001). The EMIC results refer to simulations in which climate is in equilibrium with an atmospheric CO_2 concentration of 280 ppmv. From Figures 8.17a and 8.17b, it can be seen that the simulated latitudinal distributions of the zonally averaged surface air temperature for boreal winter and boreal summer are in good agreement with observations, except at northern and southern high latitudes. Interestingly, also the GCM results exhibit a larger scatter in these regions, and they somewhat deviate from data there. Figures 8.17c and 8.17d indicate that EMICs satisfactorily reproduce the general structure of the observed zonally averaged precipitation. Here again, for most latitudes, the scatter in the EMIC results seems to be as large as the scatter in the GCM results, and both EMIC and GCM results agree with observational estimates. When these EMICs are allowed to adjust to a doubling of atmospheric CO_2 concentration, they all experience an increase in globally averaged, annual mean surface temperature and precipitation which falls by and large within the range of GCM results (Petoukhov et al., 2005).

[INSERT FIGURE 8.17 HERE]

The responses of the North Atlantic MOC to increasing atmospheric CO_2 concentration and idealised freshwater perturbations as simulated by EMICs have also been compared to those obtained by AOGCMs (Petoukhov et al., 2005; Gregory et al., 2005; Stouffer et al., 2006). These studies reveal no systematic difference in model behaviour, which gives added confidence to the use of EMICs.

In a further intercomparison, Rahmstorf et al. (2005) compared results from eleven EMICs in which the North Atlantic Ocean was subjected to a slowly varying change in freshwater input. All the models analysed show a characteristic hysteresis response of the North Atlantic MOC to freshwater forcing, which can be explained by Stommel's (1961) salt advection feedback. The width of the hysteresis curve varies between 0.2 and 0.5 Sv in the models. Major differences are found in the location of the present-day climate on the hysteresis diagram. In seven of the models, the present-day climate for standard parameter choices is found in the bi-stable regime, while in the other four models, this climate is situated in the mono-stable regime. The proximity of the present-day climate to Stommel's bifurcation point, beyond which North Atlantic Deep Water formation cannot be sustained, varies from less than 0.1 Sv to over 0.5 Sv.

A final example of EMIC intercomparison is discussed in Brovkin et al. (2006). EMICs that explicitly simulate the interactions between atmosphere, ocean and land surface were forced by a reconstruction of land cover changes during the last millennium. In response to historical deforestation of about 18×10^6 km^2, all models exhibit a decrease in globally averaged, annual mean surface temperature in the range of 0.13–0.25°C, mainly due to the increase in land surface albedo. Further experiments with the models forced by historical atmospheric CO_2 trend reveal that, for the whole last millennium, the biogeophysical cooling due to land cover changes is less pronounced than the warming induced by elevated atmospheric CO_2 level (0.27–0.62°C). During the 19th century, the cooling effect of deforestation appears to counterbalance, albeit not completely, the warming effect of increasing CO_2 concentration.

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Frequently Asked Question 8.1: How Reliable Are the Models Used to Make Projections of Future Climate Change?

There is considerable confidence that climate models provide credible quantitative estimates of future climate change, particularly at continental scales and above. This confidence comes from the foundation of the models in accepted physical principles and from their ability to reproduce observed features of current climate and past climate changes. Confidence in model estimates is higher for some climate variables (e.g., temperature) than for others (e.g., precipitation). Over several decades of development, models have consistently provided a robust and unambiguous picture of significant climate warming in response to increasing greenhouse gases.

Climate models are mathematical representations of the climate system, expressed as computer codes and run on powerful computers. One source of confidence in models comes from the fact that model fundamentals are based on established physical laws, such as conservation of mass, energy and momentum, along with a wealth of observations.

A second source of confidence comes from the ability of models to simulate important aspects of the current climate. Models are routinely and extensively assessed by comparing their simulations with observations of the atmosphere, ocean, cryosphere, and land surface. Unprecedented levels of evaluation have taken place over the last decade in the form of organised multi-model 'intercomparisons'. Models show significant and increasing skill in representing many important mean climate features, such as the large-scale distributions of atmospheric temperature, precipitation, radiation, and wind, and of oceanic temperatures, currents and sea-ice cover. Models can also simulate essential aspects of many of the patterns of climate variability observed across a range of timescales. Examples include the advance and retreat of the major monsoon systems, the seasonal shifts of temperatures, storm tracks and rain belts, and the hemispheric scale see-sawing of extra tropical surface pressures (the Northern and Southern 'annular modes'). Some climate models, or closely related variants, have also been tested by using them to predict weather and make seasonal forecasts. These models demonstrate skill in such forecasts, showing they can represent important features of the general circulation across shorter timescales, as well as aspects of seasonal and interannual variability. Models' ability to represent these and other important climate features increases our confidence that they represent the essential physical processes important for the simulation of future climate change. (Note that the limitations in climate models' ability to forecast weather beyond a few days do not limit their ability to predict long-term climate changes, as these are very different types of prediction -- see FAQ 1.2)

A third source of confidence comes from the ability of models to reproduce features of past climates and climate changes. Models have been used to simulate ancient climates, such as the warm mid-Holocene of 6000 years ago, or the last glacial maximum of 21,000 years ago (see Chapter 6). They can reproduce many features (allowing for uncertainties in reconstructing past climates) such as the magnitude and broad scale pattern of oceanic cooling during the last ice age. Models can also simulate many observed aspects of climate change over the instrumental record. One example is that the global temperature trend over the past century (shown in Figure 1) can be modeled with high skill when both human and natural factors that influence climate are included. Models also reproduce other observed changes, such as the faster increase in night time than daytime temperatures, the larger degree of warming in the Arctic, and the small, short-term global cooling (and subsequent recovery) which has followed major volcanic eruptions, such as that of Mt. Pinatubo in 1991 (see Figure). Model global temperature projections made over the last two decades have also been in overall agreement with subsequent observations over that period (Chapter 1).

[INSERT FAQ 8.1, FIGURE 1 HERE]

Nevertheless, models still show significant errors. Although these are generally greater at smaller scales, important large-scale problems also remain. For example, deficiencies remain in the simulation of tropical precipitation, the El Niño-Southern Oscillation and the Madden-Julian Oscillation (an observed variation in tropical winds and rainfall with a timescale of 30–90 days). The ultimate source of most such errors is that many important small-scale processes cannot be represented explicitly in models, and so must be included in approximate form as they interact with larger scale features. This is partly due to limitations in computing power, but also results from limitations in scientific understanding or in the availability of detailed observations of some physical processes. Significant uncertainties, in particular, are associated with the representation of clouds, and in the resulting cloud responses to climate change. As a consequence, models continue to display a substantial range of global temperature change in response to specified greenhouse gas forcing (see Chapter 10). Despite such uncertainties, however, models are unanimous in their prediction of substantial climate warming under greenhouse gas increases, and this warming is of a magnitude consistent with independent estimates derived from other sources, such as from observed climate changes and past climate reconstructions.

Since confidence in the changes projected by global models decreases at smaller scales, other techniques, such as the use of regional climate models, or downscaling methods, have been specifically developed for the study of regional and local scale climate change (see FAQ 11.1). However, as global models continue to develop, and their resolution continues to improve, they are becoming increasingly useful for investigating important smaller scale features, such as changes in extreme weather events, and further improvements in regional scale representation are expected with increased computing power. Models are also becoming more comprehensive in their treatment of the climate system, thus explicitly representing more physical and biophysical processes and interactions considered potentially important for climate change, particularly at longer timescales. Examples are the recent inclusion of plant responses, ocean biological and chemical interactions, and ice sheet dynamics in some global climate models.

In summary, confidence in models comes from their physical basis, and their skill in representing observed climate and past climate changes. Models have proven to be extremely important tools for simulating and understanding climate, and there is considerable confidence that they are able to provide credible quantitative estimates of future climate change, particularly at larger scales. Models continue to have significant limitations, such as in their representation of clouds, which lead to uncertainties in the magnitude and timing, as well as regional details, of predicted climate change. Nevertheless, over several decades of model development, they have consistently provided a robust and unambiguous picture of significant climate warming in response to increasing greenhouse gases.

Tables

Table 8.1. Table of Selected Model Features. Salient features of the AOGCMs participating in the multi-model dataset at PCMDI (MMD) are listed by IPCC ID along with the calendar year ("vintage") of the first publication of results from each model. Also listed are the respective sponsoring institutions, the pressure at the top of the atmospheric model, the horizontal and vertical resolution of the model atmosphere and ocean models, the pressure of the atmospheric top, as well as the oceanic vertical coordinate (depth or density) and upper boundary condition (free surface or rigid lid). Also listed are the characteristics of sea ice dynamics/structure (e.g., rheology vs. "free drift" assumption and inclusion of ice leads), and whether adjustments of surface momentum, heat, or freshwater fluxes are applied in coupling the atmosphere, ocean, and sea ice components. Land features such as the representation of soil moisture (single-layer "bucket" vs. multi-layered scheme) and the presence of a vegetation canopy or a river routing scheme also are noted. Relevant references describing details of these aspects of the models also are cited.

Model ID, Vintage	Sponsor(s), Country	Atmosphere Top Resolution References	Ocean Resolution Z Coord., Top BC References	Sea Ice Dynamics, Leads References	Coupling Flux Adjustments References	Land Soil, Plants, Routing References
1: BCC-CM1, 2005	Beijing Climate Center, China	top = 25 hPa T63 (1.9°×1.9°)L16 Dong et al., 2000 CSMD, 2005 Xu et al., 2005	1.9° × 1.9° L30 depth, free surface Jin et al., 1999	no rheology or leads Xu et al., 2005	heat, momentum Yu & Zhang, 2000 CSMD, 2005	layers, canopy, routing CSMD, 2005
2: BCCR-BCM2.0, 2005	Bjerknes Centre for Climate Research, Norway	top = 10 hPa T63(1.9° × 1.9°)L31 Déqué et al., 1994	0.5–1.5° × 1.5° L35 density, free surface Bleck et al., 1992	rheology, leads Hibler, 1979, Harder, 1996	no adjustments Furevik et al., 2003	layers,canopy,routing Mahfouf et al., 1995 Douville et al., 1995 Oki & Sud, 1998
3: CCSM3, 2005	National Center for Atmospheric Research, USA	top = 2.2 hPa T85(1.4° x 1.4°)L26 Collins et al,. 2004	0.3–1° × 1° L40 depth, free surface Smith & Gent, 2002	rheology, leads Briegleb et al., 2004	no adjustments Collins et al., 2006	layers, canopy, routing Oleson et al., 2004 Branstetter, 2001
4: CGCM3.1(T47), 2005		top = 1 hPa T47(~2.8° x 2.8°)L31 McFarlane et al., 1992; Flato, 2005	1.9° × 1.9° L29 depth, rigid lid Pacanowski et al., 1993	rheology, leads Hibler, 1979 Flato & Hibler, 1992	heat, fresh water Flato, 2005	layers, canopy, routing Verseghy et al., 1993
5: CGCM3.1(T63), 2005	Canadian Centre for Climate Modeling & Analysis, Canada	top = 1 hPa T63(~1.9° x 1.9°)L31 McFarlane et al., 1992; Flato 2005	0.9° × 1.4° L29 depth, rigid lid Flato & Boer, 2001 Kim et al., 2002	rheology, leads Hibler, 1979 Flato & Hibler, 1992	heat, fresh water Flato, 2005	layers, canopy, routing Verseghy et al., 1993
6: CNRM-CM3, 2004	Météo-France/Centre National de Recherches Météorologiques, France	top = 0.05 hPa T63(~1.9° x 1.9°)L45 Déqué et al., 1994	0.5–2° × 2° L31 depth, rigid lid Madec et al., 1998	rheology, leads Hunke-Dukowicz, 1997; Salas-Mélia, 2002	no adjustments Terray et al., 1998	layers, canopy,routing Mahfouf et al., 1995 Douville et al., 1995; Oki & Sud, 1998

Model ID, Vintage	Sponsor(s), Country	Atmosphere Top Resolution References	Ocean Resolution Z Coord., Top BC References	Sea Ice Dynamics, Leads References	Coupling Flux Adjustments References	Land Soil, Plants, Routing References
7: CSIRO-MK3.0, 2001	CSIRO Atmospheric Research, Australia	top = 4.5 hPa T63(~1.9° x 1.9°)L18 Gordon et al., 2002	0.8° x 1.9° L31 depth, rigid lid Gordon et al., 2002	rheology, leads O'Farrell, 1998	no adjustments Gordon et al., 2002	layers, canopy Gordon et al., 2002
8: ECHAM5/MPI-OM, 2005	Max Planck Institute for Meteorology, Germany	top = 10 hPa T63(~1.9° x 1.9°)L31 Roeckner et al., 2003	1.5° x 1.5° L40 depth, free surface Marsland et al., 2003	rheology, leads Hibler, 1979, Semtner, 1976	no adjustments Jungclaus et al., 2005	bucket, canopy, routing Hagemann, 2002 Hagemann & Dümenil–Gates, 2001
9: ECHO-G, 1999	Meteorological Institute of the University of Bonn, Meteorologi-cal Research Institute of KMA, and Model & Data Group, Germany/Korea	top = 10 hPa T30 (~3.9° x 3.9°)L19 Roeckner et al., 1996	0.5–2.8° x 2.8° L20 depth, free surface Wolff et al., 1997	rheology, leads Wolff et al., 1997	heat, freshwater Min et al., 2005	bucket, canopy, routing Roeckner et al., 1996 Dümenil & Todini, 1992
10: FGOALS-g1.0, 2004	LASG/Institute of Atmospheric Physics, China	top = 2.2 hPa T42(~2.8° x 2.8°)L26 Wang et al., 2004	1.0° x 1.0° L16 eta, free surface Jin et al., 1999; Liu et al., 2004	rheology, leads Briegleb et al., 2004	no adjustments Yu et al. 2002, 2004	layers, canopy,routing Bonan et al., 2002
11: GDFL-CM2.0, 2005	U.S. Dept. of Commerce/NOAA/ Geophysical Fluid Dynamics Laboratory, USA	top = 3 hPa 2.0° x 2.5° L24 GFDL GAMDT, 2004	0.3–1.0° x 1.0° depth, free surface Gnanadesikan et al., 2004	rheology, leads Winton, 2000; Delworth et al., 2006	no adjustments Delworth et al., 2006	bucket, canopy, routing Milly & Shmakin, 2002; GFDL GAMDT, 2004
12: GDFL-CM2.1, 2005		top = 3 hPa 2.0° x 2.5° L24 GFDL GAMDT, 2004 with semi-Lagrangian transports	0.3–1.0° x 1.0° depth, free surface Gnanadesikan et al., 2004	rheology, leads Winton, 2000; Delworth et al., 2006	no adjustments Delworth et al., 2006	bucket, canopy, routing Milly & Shmakin, 2002; GFDL GAMDT, 2004
13: GISS-AOM, 2004		top = 10 hPa 3° x 4° L12 Russell et al., 1995; Russell, 2005	3 x 4° L16 mass/area, free sfc. Russell et al., 1995; Russell, 2005	rheology, leads Flato & Hibler, 1992 Russell, 2005	no adjustments Russell, 2005	layers, canopy, routing Abramopoulos et al., 1988; Miller et al., 1994
14: GISS-EH, 2004	NASA/Goddard Institute for Space Studies, USA	top = 0.1 hPa 4° x 5° L20 Schmidt et al., 2006	2° x 2° L16 density, free surface Bleck, 2002	rheology, leads Liu et al., 2003; Schmidt et al., 2004	no adjustments Schmidt et al., 2006	layers, canopy, routing Friend & Kiang, 2005

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Model ID, Vintage	Sponsor(s), Country	Atmosphere Top Resolution References	Ocean Resolution Z Coord., Top BC References	Sea Ice Dynamics, Leads References	Coupling Flux Adjustments References	Land Soil, Plants, Routing References
15: GISS-ER, 2004	NASA/Goddard Institute for Space Studies, USA	top = 0.1 hPa 4° x 5° L20 Schmidt et al., 2006	4° x 5° L13 mass/area, free sfc. Russell et al., 1995	rheology, leads Liu et al., 2003; Schmidt et al., 2004	no adjustments Schmidt et al., 2006	layers, canopy, routing Friend & Kiang, 2005
16: INM-CM3.0, 2004	Institute for Numerical Mathematics, Russia	top = 10 hPa 4º x 5º L21 Alekseev et al., 1998; Galin et al., 2003	2º x 2.5º L33 sigma, rigid lid Diansky et al., 2002	no rheology or leads Diansky et al., 2002	regional freshwater Diansky & Volodin, 2002; Volodin & Diansky, 2004	layers, canopy, no routing Alekseev et al., 1998; Volodin & Lykosoff, 1998
17: IPSL-CM4, 2005	Institut Pierre Simon Laplace, France	top = 4 hPa 2.5º x 3.75º L19 Hourdin et al., 2006	2º x 2º L31 depth, free surface Madec et al., 1998	rheology, leads Fichefet & Morales Maqueda, 1997 Goosse & Fichefet, 1999	no adjustments Marti et al., 2005	layers, canopy, routing Krinner et al., 2005
18: MIROC3.2(hires), 2004	Center for Climate System Research (University of Tokyo), National Institute for Environmental Studies, and Frontier Research Center for Global Change (JAMSTEC), Japan	top = 40 km T106(~1.1° x 1.1°)L56 K-1 Developers, 2004	0.2º x 0.3º L47 sigma/depth, free surface K-1 Developers, 2004	rheology, leads K-1 Developers, 2004	no adjustments K-1 Developers, 2004	layers, canopy, routing K-1 Developers, 2004 Oki & Sud, 1998
19: MIROC3.2(medres), 2004		top = 30 km T42(~2.8º x 2.8º)L20 K-1 Developers, 2004	0.5–1.4º x 1.4º L43 sigma/depth, free surface K-1 Developers, 2004	rheology, leads K-1 Developers, 2004	no adjustments K-1 Developers, 2004	layers, canopy, routing K-1 Developers, 2004 Oki & Sud, 1998
20: MRI-CGCM2.3.2, 2003	Meteorological Research Institute, Japan	top = 0.4 hPa T42(~2.8º x 2.8º)L30 Shibata et al., 1999	0.5–2.0º x 2.5º L23 depth, rigid lid Yukimoto et al. 2001	free drift, leads Mellor & Kantha, 1989	heat, freshwater, momentum (12S–12N) Yukimoto et al., 2001; Yukimoto & Noda, 2003	layers,canopy, routing Sellers et al., 1986, Sato et al., 1989
21: PCM, 1998	National Center for Atmospheric Research, USA	top = 2.2 hPa T42(~2.8º x 2.8º)L26 Kiehl et al., 1998	0.5–0.7º x 1.1º L40 depth, free surface Maltrud et al., 1998	rheology, leads Hunke & Dukowicz 1997, 2003 Zhang et al., 1999	no adjustments Washington et al., 2000	layers, canopy, no routing Bonan, 1998
22: UKMO-HadCM3, 1997	Hadley Centre for Climate	top = 5 hPa 2.5º x 3.8º L19 Pope et al., 2000	1.5º x 1.5º L20 depth, rigid lid Gordon et al., 2000	free drift, leads Cattle & Crossley, 1995	no adjustments Gordon et al., 2000	layers,canopy,routing Cox et al., 1999

23: UKMO-HadGEM, 2004	Prediction and Research/Met Office, UK	top = 39.2 km ~1.3º x 1.9º L38 Martin et al., 2004	0.3–1.0º x 1.0º L40 depth, free surface Roberts, 2004	rheology, leads Hunke & Dukowicz, 1997; Semtner, 1976; Lipscomb, 2001	no adjustments Johns et al., 2006	layers, canopy, routing Essery et al., 2001; Oki & Sud, 1998

Table 8.2. Climate sensitivity estimates from the AOGCMs assessed in this report. Transient climate response (TCR) and equilibrium climate sensitivity (ECS) have been calculated by the modelling groups (using atmosphere models coupled to slab ocean for equilibrium climate sensitivity), except those marked [a], which were calculated from simulations in the multi-model dataset at PCMDI. The ocean heat uptake efficiency ($W\ m^{-2}\ K^{-1}$), discussed in Chapter 10, may be roughly estimated as $F_{2\times}(TCR^{-1} - ECS^{-1})$, where $F_{2\times}$ is the radiative forcing for doubled CO_2 concentration (see Supplementary Material, Table 8.SM.1)

AOGCM	Equilibrium climate sensitivity (K)	Transient climate response (K)
1: BCC-CM1, China	n/a	n/a
2: BCCR-BCM2.0, Norway	n/a	n/a
3: CCSM3, USA	2.7	1.5
4: CGCM3.1(T47), Canada	3.4	1.9 [a]
5: CGCM3.1(T63), Canada	3.4 [a]	n/a
6: CNRM-CM3, France	n/a	1.6
7: CSIRO-Mk3.0, Australia	3.1	1.4
8: ECHAM5/MPI-OM, Germany	3.4	2.2
9: ECHO-G, Germany/Korea	3.2	1.7
10: FGOALS-g1.0, China	2.3 [a]	1.2 [a]
11: GFDL-CM2.0, USA	2.9	1.6
12: GFDL-CM2.1, USA	3.4	1.5
13: GISS-AOM, USA	n/a	n/a
14: GISS-EH, USA	2.7	1.6
15: GISS-ER, USA	2.7	1.5
16: INM-CM3.0, Russia	2.1	1.6
17: IPSL-CM4, France	4.4	2.1
18: MIROC3.2(hires), Japan	4.3	2.6
19: MIROC3.2(medres), Japan	4.0	2.1
20: MRI-CGCM2.3.2, Japan	3.2	2.2
21: PCM, USA	2.1	1.3
22: UKMO-HadCM3, UK	3.3	2.0
23: UKMO-HadGEM1, UK	4.4	1.9

Table 8.3. Description of the EMICs used in Chapter 10. The naming convention for the models is as agreed by all modelling groups involved.

NAME	ATMOSPHERE	OCEAN	SEA ICE	COUPLING / FLUX ADJUSTMENTS	LAND SURFACE	BIOSPHERE	ICE SHEETS
E1: BERN2.5CC (Plattner et al., 2001; Joos et al., 2001)	EMBM, 1-D(φ), NCL, 7.5°–15° (Schmittner and Stocker, 1999)	FG with parameterised zonal pressure gradient, 2-D(φ, z), 3 basins, RL, ISO, MESO, 7.5°–15°, L14 (Wright and Stocker, 1992)	0-LT, 2-LIT (Wright and Stocker, 1993)	PM, NH, NW (Stocker et al., 1992; Schmittner and Stocker, 1999)	NST, NSM (Schmittner and Stocker, 1999)	BO (Marchal et al., 1998), BT (Sitch et al., 2003; Gerber et al., 2003), BV (Sitch et al., 2003; Gerber et al., 2003)	-
E2: C-GOLDSTEIN (Edwards and Marsh, 2005)	EMBM, 2-D(φ, λ), NCL, 5° × 10° (Edwards and Marsh, 2005)	FG, 3-D, RL, ISO, MESO, 5° × 10°, L8 (Edwards and Marsh, 2005)	0-LT, DOC, 2-LIT (Edwards and Marsh, 2005)	GM, NH, RW (Edwards and Marsh, 2005)	NST, NSM, RIV (Edwards and Marsh, 2005)	-	-
E3: CLIMBER-2 (Petoukhov et al., 2000)	SD, 3-D, CRAD, ICL, 10° × 51°, L10 (Petoukhov et al., 2000)	FG with parameterised zonal pressure gradient, 2-D(φ, z), 3 basins, RL, 2.5°, L21 (Wright and Stocker, 1992)	0-LT, DOC, 2-LIT (Petoukhov et al., 2000)	NM, NH, NW (Petoukhov et al., 2000)	1-LST, CSM, RIV (Petoukhov et al., 2000)	BO (Brovkin et al., 2002), BT (Brovkin et al., 2002), BV (Brovkin et al., 2002)	TM, 3-D, 0.75° × 1.5°, L20* (Calov et al., 2005)
E4: CLIMBER-3α (Montoya et al., 2005)	SD, 3-D, CRAD, ICL, 7.5° × 22.5°, L10 (Petoukhov et al., 2000)	PE, 3-D, FS, ISO, MESO, TCS, DC*, 3.75° × 3.75°, L24 (Montoya et al., 2005)	2-LT, R, 2-LIT (Fichefet and Morales Maqueda, 1997)	AM, NH, RW (Montoya et al., 2005)	1-LST, CSM, RIV (Petoukhov et al., 2000)	BO* (Six and Maier-Reimer, 1996), BT* (Brovkin et al., 2002), BV* (Brovkin et al., 2002)	-
E5: LOVECLIM (Driesschaert, 2005)	QG, 3-D, LRAD, NCL, T21 (5.6° × 5.6°), L3 (Opsteegh et al., 1998)	PE, 3-D, FS, ISO, MESO, TCS, DC, 3° × 3°, L30 (Goosse and Fichefet, 1999)	3-LT, R, 2-LIT (Fichefet and Morales Maqueda, 1997)	NM, NH, RW (Driesschaert., 2005)	1-LST, BSM, RIV (Opsteegh et al., 1998)	BO (Mouchet and François, 1996), BT (Brovkin et al., 2002), BV (Brovkin et al., 2002)	TM, 3-D, 10 km × 10 km, L30 (Huybrechts, 2002)
E6: MIT-IGSM2.3 (Sokolov et al., 2005)	SD, 2-D(φ, z), CRAD, ICL, 4°, L11 (Sokolov and Stone, 1998) CHEM* (Mayer et	PE, 3-D, FS, ISO, MESO, 4° × 4°, L15 (Marshall et al., 1997)	3-LT, 2-LIT (Winton, 2000)	AM, GH, GW (Sokolov et al., 2005)	10-LST, CSM (Bonan et al., 2002)	BO (Parekh et al., 2005), BT (Felzer et al., 2005), BV* (Felzer et al., 2005)	-

	al., 2000)						
E7: MOBIDIC (Crucifix et al., 2002)	QG, 2-D(φ, z), CRAD, NCL, 5°, L2 (Gallée et al., 1991)	PE with parameterised zonal pressure gradient, 2-D(φ, z), 3 basins, RL, DC, 5°, L15 (Hovine and Fichefet, 1994)	0-LT, PD, 2-LIT (Crucifix et al., 2002)	NM, NH, NW (Crucifix et al., 2002)	1-LST, BSM (Gallée et al., 1991)	BO* (Crucifix, 2005), BT* (Brovkin et al., 2002), BV (Brovkin et al., 2002)	M, 1-D(φ), 0.5° (Crucifix and Berger, 2002)
E8: UVIC (Weaver et al., 2001)	DEMBM, 2-D(φ, λ), NCL, 1.8° × 3.6° (Weaver et al., 2001)	PE, 3-D, RG, ISO, MESO, 1.8° × 3.6° (Weaver et al., 2001)	0-LT, R, 2-LIT (Weaver et al., 2001)	AM, NH, NW (Weaver et al., 2001)	1-LST, CSM, RIV (Meissner et al., 2003)	BO (Weaver et al., 2001), BT (Cox, 2001), BV (Cox, 2001)	M, 2-D(φ, λ), 1.8° × 3.6°* (Weaver et al., 2001)

Notes:

Atmosphere: EMBM = energy-moisture balance model; DEMBM = energy-moisture balance model including some dynamics; SD = statistical-dynamical model; QG = quasi-geostrophic model; 1-D(φ) = zonally and vertically averaged; 2-D(φ, λ) = vertically averaged; 2-D(φ, z) = zonally averaged; 3-D = three-dimensional; LRAD = linearised radiation scheme; CRAD = comprehensive radiation scheme; NCL = non-interactive cloudiness; ICL = interactive cloudiness; CHEM = chemistry module; horizontal and vertical resolutions: the horizontal resolution is expressed either as degrees latitude × longitude or as spectral truncation with a rough translation to degrees latitude × longitude; the vertical resolution is expressed as "Lm", where m is the number of vertical levels.

Ocean: FG = frictional geostrophic model; PE = primitive equation model; 2-D(φ, z) = zonally averaged; 3-D = three-dimensional; RL = rigid lid; FS = free surface; ISO = isopycnal diffusion; MESO = parameterisation of the effect of mesoscale eddies on tracer distribution; TCS = complex turbulence closure scheme; DC = parameterisation of density-driven downsloping currents; horizontal and vertical resolutions: the horizontal resolution is expressed as degrees latitude × longitude; the vertical resolution is expressed as "Lm", where m is the number of vertical levels.

Sea ice: n-LT = n-layer thermodynamic scheme; PD = prescribed drift; DOC = drift with oceanic currents; R = viscous-plastic or elastic-viscous-plastic rheology; 2-LIT = two-level ice thickness distribution (level ice and leads).

Coupling / flux adjustments: PM = prescribed momentum flux; GM = global momentum flux adjustment; AM = momentum flux anomalies relative to the control run are computed and added to climatological data; NM = no momentum flux adjustment; GH = global heat flux adjustment; NH = no heat flux adjustment; GW = global freshwater flux adjustment; RW = regional freshwater flux adjustment; NW = no freshwater flux adjustment.

Land surface: NST = no explicit computation of soil temperature; n-LST = n-layer soil temperature scheme; NSM = no moisture storage in soil; BSM = bucket model for soil moisture; CSM = complex model for soil moisture; RIV = river routing scheme.

Biosphere: BO = model of oceanic carbon dynamics; BT = model of terrestrial carbon dynamics; BV = dynamical vegetation model.

Ice sheets: TM = thermomechanical model; M = mechanical model (isothermal); 1-D(φ) = vertically averaged with east-west parabolic profile; 2-D(φ, λ) = vertically averaged; 3-D = three-dimensional; horizontal and vertical resolutions: the horizontal resolution is expressed either as degrees latitude × longitude or kilometres × kilometres; the vertical resolution is expressed as "Lm", where m is the number of vertical levels.

An asterisk after a component or parameterisation means that this component or parameterisation was not activated in the experiments discussed in Chapter 10.