Chapter 8: Climate Models and Their Evaluation

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Executive Summary

The goal of this chapter is to assess the capacity of the global climate models used elsewhere in this report, for projecting future climate change. In the TAR, it was concluded that AOGCMs could provide useful projections of future climate. Our confidence in this conclusion has been enhanced via a suite of advances since the TAR.

There is considerable confidence that models are reliable enough to provide useful projections of future climate change, particularly at larger scales. 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. In this summary we focus on areas of progress since the TAR.

Highlights since the TAR include:

- There have been ongoing improvements to resolution, numerics and parametrisations, and more processes (e.g., interactive aerosols) have been included in models.
- Fewer models use flux adjustments. Most AR4 AOGCMs do not use flux adjustments to maintain a stable climate. The uncertainty associated with the use of those adjustments is therefore decreasing. However, in spite of various model improvements, climate drift remains an issue in most AOGCM control simulations.
- There have been improvements in the simulation of many aspects of present climate. The fact that this improvement has continued alongside the decreasing use of flux adjustments points to the progress made in the development of the numerical algorithms and physical parameterizations and in the increased resolution in these models. At least some of these improvements can be traced to the above improvements in formulation.
- Model simulation of modes of climate variability has been more thoroughly assessed than previously. Many aspects of observed variability are well simulated, but deficiencies still exist for some of the major modes (e.g., ENSO). Models have skill in simulating extremes of temperature and wind, though extreme precipitation remains more elusive.
- Substantial progress has been made in understanding the differences between different models' estimates of equilibrium climate sensitivity. Cloud feedbacks have been confirmed as a primary source of inter-model differences, with tropical low cloud the largest contributor. New evidence strongly supports a combined water vapour-lapse rate feedback of around the strength found in GCMs. The magnitude of cryospheric feedbacks remains uncertain, leading to a range of possible climate response at mid-to-high latitudes.
- Progress has been made in understanding the wide range of modelled response of the Atlantic meridional overturning circulation (MOC) under greenhouse gas forcing. However the processes involved are not always amenable to testing against observation, so a wide range of possible MOC responses remains. Systematic biases have been noted in most models' simulation of the Southern Ocean; since the Southern Ocean is an important region for ocean heat uptake this results in some uncertainty in transient climate response.
- The advent of large ensembles of climate models has enhanced understanding of how particular observational tests may constrain climate projections. However the goal of a proven model metric that can narrow the range of plausible climate projections has not yet been achieved.
- A small number of climate models have been tested, and show useful skill, in forecasting on timescales from daily to decadal, when initialised with observed conditions. Such successful forecasts increase confidence in the models' representation of some of the key processes for longer term climate projection.
- Carbon cycle feedbacks have been included in a few climate models, but not as yet in the main climate projections in Chapter 10. The magnitude of these feedbacks over the 21st Century remains uncertain, but they are likely to be (a) positive and (b) relatively small over the next few decades.

Developments in model formulation

Climate models have changed significantly since the TAR, and so have the methods of model evaluation.. Changes include:

- Improved numerical schemes and parameterizations
- Higher spatial and temporal resolution
- A more comprehensive range of components included in some climate models
- Better simulations of some aspects of the current climate, including specific modes of variability
- Several new ways of testing climate models
- Community-wide scrutiny of the model evaluation process
- Improved computational strategies, e.g., larger ensembles of simulations, and multi-model ensembles
- Improved understanding of the processes responsible for range of model results.

Improvements in atmospheric models include reformulated dynamics (e.g., semi-Lagrangian advection), and increased horizontal and vertical resolution. Interactive aerosol component modules have been incorporated into some atmospheric models, and through these the direct and the indirect effect of aerosols are now more widely simulated. In at least one case, improvements to a model's boundary-layer parametrisation are believed to have played a key role in improving the model's marine stratocumulus simulation. The parametrisation had been tested independently of the GCM through a community-based project.

Significant developments have occurred in the representation of terrestrial processes, with most models now representing these processes as well as the best model did in the TAR. Individual components continue to be improved via a systematic evaluation against observations and against other models. The representation of terrestrial processes in climate models is generally good enough that we have no reason to think that they cannot capture the main processes that significantly affect large-scale climate over the next few decades. However, there is emerging evidence that longer-term changes in terrestrial carbon storage, in the vegetation and soils, can generate major changes in the terrestrial carbon sink. Several climate models have now included these processes, but they are not yet used in the body of climate projections presented in Chapter 10. The limited evidence indicated that the magnitude of the carbon cycle feedback on climate varies widely, from small to significantly positive, on timescales of 50–100 years, depending substantially on the model's climate sensitivity.

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

The progress in developing AOGCM cryospheric components is clearest for sea ice. Almost all state-of-the-art AOGCMs now include fairly elaborate sea-ice dynamics ranging from the cavitating fluid to the most advanced elastic-viscous-plastic rheology. Only few models still use either motionless ice cover, or that advected with ocean currents. Some AOGCMs now include several sea-ice thickness categories and relatively advanced thermodynamics.

AOGCM parameterizations of terrestrial snow processes vary from rather simplistic to those accounting for the impact of overlying vegetation, snow ripening, re-distribution by wind, and the impact of dust on albedo. Efforts that have evaluated this component suggest that surface tiling and sub-grid scale heterogeneity of snow are key to capturing observations of seasonal snow cover. Representation of other terrestrial cryosphere components in AOGCMs (ice-sheets, permafrost, etc.) lags behind the state of the art in their stand-alone modeling. For example, only few AOGCMs include ice sheet dynamics, and none of the AOGCM versions evaluated in this chapter represents ice sheets other than in the most simplistic form.

Most AR4 AOGCMs do not use flux adjustments to maintain a stable climate. The uncertainty associated with the use of those adjustments is therefore decreasing. However, in spite of various model improvements, climate drift remains an issue in most AOGCM control simulations.

Developments in model climate simulation

There has been a noticeable improvement in the ability of some models to simulate marine subtropical stratocumulus clouds, which are important for the simulation of sea surface temperature.

Development of the new AR4 models has not led to qualitative changes since the TAR in projections of future ocean changes. This increases confidence in the robustness of those projections. However the importance of the Southern Ocean in determining the rate of ocean heat uptake has been identified, and some common model biases in that region result in some uncertainty in transient climate response. Overall, model simulation of ocean water mass structure, overturning circulation and heat transport has improved since the TAR. It is likely that at least part of the improvement is due to the improvements in formulation mentioned above. Specifically, the over-thick thermocline and deficient Atlantic overturning and heat transport, common in the TAR, are substantially improved in many models.

In spite of the notable progress in developing AOGCM sea ice components and an improved ability of some models to better capture key features of sea-ice geographical distribution and seasonality, since TAR the AOGCMs as a class have demonstrated only a modest improvement in simulations of the current sea-ice climate. The relatively slow progress may be at least partly explained by the fact that improving sea ice distribution depends also on improvements in both atmospheric and oceanic general circulation simulations.

Since the TAR there has been progress in the representation of large-scale variability over a wide range of time-scales in coupled GCMs used for climate projections. Coupled GCMs capture the dominant extratropical patterns of variability known as the Northern and Southern Annular Modes (NAM and SAM), the Pacific Decadal Oscillation (PDO) and the Pacific-North American (PNA) and Cold Ocean-Warm Land (COWL) Patterns. Coupled GCMs simulate Atlantic multidecadal variability although the relative roles of high and low latitude processes appear to differ from model to model. In the tropics, obtaining a completely accurate representation of the El Niño-Southern Oscillation (ENSO) and the Madden-Julian Oscillation (MJO) with coupled GCMs continues to present a challenge. Developments in model formulation since the TAR have generally led to improvements in the amplitude, structure and time-scale of these modes, yet systematic errors persist.

GCMs are showing good skill in simulating extreme temperatures and the number of frost days but their ability to simulate extreme precipitation is still poor. These models tend to produce too many days with weak precipitation and too few days with high precipitation.

Given the large computing resources required by coupled GCMs, Earth system models of intermediate complexity (EMICs) are widely utilised to study past and future climate changes. Since the TAR, a great deal of effort has been devoted to the evaluation of those models through organised model intercomparisons. These exercises have revealed that, at large scales, EMIC results compare reasonably well with observational data and coupled GCM results. This gives confidence to the use of these models to understand important processes and their interactions within the climate system and to explore uncertainties in long-term climate change projections. However, because of their reduced resolution and simplified representation of some physical processes, it would not be sensible to apply an EMIC to study small-scale processes.

Developments in analysis methods

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

Water vapour feedback remains the most important positive feedback in models. Although there is a spread among models in the magnitude of their water vapour and lapse rate feedbacks, impact on climate sensitivity is reduced by anti-correlation between them. Several new studies indicate that current climate models simulate the response of lower and upper tropospheric relative humidity seasonal and interannual variability, volcanic induced cooling and climate trends, in a way consistent with observations (within the range of observational uncertainties). Furthermore there is no substantial evidence to suggest that the nearly unchanged relative humidity response predicted by general circulation models under climate change constitutes a model artifact. Taken together, observational and modelling evidence strongly favour a combined water vapour-lapse rate feedback of around the strength found in GCMs.

On the other hand, recent studies reaffirm that the spread of climate sensitivity estimates among models primarily arises from intermodel differences in cloud feedbacks. The shortwave response to climate change of tropical boundary-layer clouds, and to a lesser extent mid-level clouds, constitutes the largest contributor to intermodel differences in global cloud feedbacks. The relatively poor simulation of these clouds in the present climate, especially in the eastern tropical oceans, is a reason for some concern. The response to global warming of upper-level clouds is also a significant source of uncertainty since current models exhibit substantial biases in the simulation of deep convective clouds. Based on the observational tests currently used to evaluate components of cloud feedbacks, each climate model exhibits particular strengths and weaknesses, and it is not yet possible to determine which model estimate of the climate change cloud feedbacks is the most reliable.

While evaluating cryospheric feedbacks in recent years has been marked by a certain progress, substantial uncertainty remains as to their magnitudes, and their representation in AOGCMs. This is one factor contributing to a spread of modelled climate responses in high latitudes. On the global scale the surface albedo feedback is positive in all the models, with a spread among current models much smaller than that of cloud feedbacks. 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. Scarcity of observations in polar regions (e.g., of sea ice thickness) also hampers evaluation. However, new techniques allowing for estimating the sea-ice and land-snow albedo feedbacks have been developed and applied to climate models. In particular, it has been suggested that the performance of an AOGCM in reproducing the observed seasonal cycle of the land snow cover (especially the springtime melt) may constitute an indirect evaluation of the snow-albedo feedback simulated by this model in climate change.

Systematic model comparison studies have helped to establish the key processes that are responsible for variations between models in the response of the ocean to climate change (especially ocean heat uptake and thermohaline circulation changes). The importance of local and non-local feedbacks from hydrological cycle changes onto the meridional overturning circulation has been established in many models. At present not all of these feedbacks are well constrained by observations, resulting in some remaining uncertainty in ocean response.

A few climate models have been tested for (and shown) skill in initial value prediction, on timescales from weather forecasting (a few days) to decadal climate variations. The fact that, given appropriate initialisation data, these models have hindcast skill increases confidence that they are representing some of the key processes and teleconnections in the climate system. On the decadal timescale, much of the predictability appears to come from climate forcing factors (e.g., increasing greenhouse gases), suggesting that forcing changes have played a key role in recent climate variations.

What does model evaluation tell us about the reliability of climate projections?

At present, it is not possible to define a robust 'model metric' which measures the reliability of a model's climate projections. However a few studies based on large ensembles of climate model integrations have shown that available observational tests potentially have value in constraining climate sensitivity. This provides support for the idea of using observations to generate a quantitative likelihood weighting of different models, allowing a more formal probabilistic framing of climate projections. However, the relationship between particular model-observation errors and biases in climate projections will depend on model-based studies. Further, the prior choice of candidate observations is likely to be based on subjective or pragmatic considerations. Detailed analysis of processes and feedbacks, as discussed above, may help make such choices less subjective. Such analyses suggest that, among others, representation of upper tropospheric humidity, tropical low clouds, and sea ice and snow cover, and their response to perturbations, should be important factors for metrics focused on model sensitivity.

While the above procedures may in future shed more quantitative light on the reliability of climate projections, at present model evaluation activities are based on the judgement of the diverse community of climate scientists, which encapsulates our present scientific understanding of the climate system. This process forms the basis for our developing confidence in the value and reliability of climate model projections.

8.11 Introduction and Philosophy

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, and this section provides a context for those studies and a guide to direct the reader to the appropriate chapters.

The models have changed significantly since the TAR, and so have the methods of model evaluation.. Changes include:

- Improved numerical schemes and parameterizations
- A more comprehensive range of processes included in some climate models
- Better simulations of some aspects of the current climate, including specific modes of variability
- Several new ways of testing climate models
- Community-wide scrutiny of the model evaluation process
- Improved experimental designs, e.g., larger ensembles of simulations, and multi-model ensembles
- Improved understanding of the processes responsible for range of model results.

This chapter touches on all of those points.

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 sceptically. 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 try to 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 outside the framework of the complete model. Here we can make an analogy with the testing of a new aircraft. Flight tests are needed to evaluate the entire aircraft as a system, but component tests are also essential.

Component-level evaluation of climate models is widely practiced now. Numerical methods are tested in standardized test cases, 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 (see, 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. Studies can be divided into three categories: simulation of the present climate (Chapter 8), simulation of the instrumental record (Chapter 9), and simulation of paleo-climate (Chapter 6).

Simulation of the present climate is often taken to mean evaluation of aspects of model ‘control runs’ (with fixed atmospheric constituents) against contemporary climate observations. Control runs vary in their in their specification of greenhouse gas and other concentrations (e.g., preindustrial or present day values; see Table 4.1), and present climate is not in equilibrium with present forcing. Therefore one would not expect even a 'perfect model' to provide perfect agreement with observations of climate over recent decades. Nonetheless a relatively large number of observations is available to define some form of 'mean climate' (including variability and extremes) over recent decades, and therefore this forms an important test of models (Sections 8.3 to 8.5).

How far does contemporary mean climate constrain future climate projections? And how accurately do we need to model a particular contemporary climate variable (e.g., the mean seasonal cycle of surface temperature) in order to model future climate change to a given accuracy? Scientific assessment of this question is still at an early stage, but two approaches are possible. The first is to use an analysis of the processes generating climate change in model simulations (e.g., Sections 8.6, 8.7) to provide insight into which aspects of the 'mean climate state' are important; for example analysis of the sea ice – albedo feedback (Section 8.6.3.4) suggests that accurate simulation of mean sea ice fields may be of moderate importance for global climate sensitivity, and critical in determining high latitude sensitivity. The second approach is to use the emerging multi-model or 'perturbed physics' ensembles to make a 'perfect model' study of sensitivity of climate response to particular observational constraints. For example Murphy et al. (2004), Knutti and Meehl (2005) and Piani et al. (2005) show that using specific observational constraints to weight members in a perturbed physics ensemble gives tighter constraints on the ensemble distribution of climate sensitivity than if the observations are not used. On the other hand Hargreaves et al. (2004) generate an ensemble of Earth System Models of Intermediate Complexity (EMICs) that all give good simulations of present-day mean ocean temperature and salinity and atmospheric surface temperature and humidity, but find that these observational constraints alone do not give a strong constraint on the future behaviour of the ocean thermohaline circulation. All the above studies are subject to two restrictions: (i) they are dependent on the structure of the particular model or ensemble used, so conclusions may be sensitive to the inclusion of a particular process or feedback which is absent in all the driving models, (ii) a prior choice of observational constraints is required, and this may be to a large extent subjective. Therefore we are some way from a robust 'model metric' for likelihood weighting of different models; but these early results do suggest that the observational tests currently available do have value in constraining climate projections. Further useful constraints come from models' ability to simulate past climate (Chapters 6 and 9).

In comparing 'present mean climate' in models against observations, certain practical decisions are needed. For example, is it more appropriate to consider 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 that includes historical variations in forcing? Such decisions are made by individual researchers, dependent on the particular problem being studied and on resource constraints on what runs can be done. Differences between model and observations that 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

should be considered insignificant, and while space does not allow us to discuss the above issues in detail for each climate variable, they are taken into account in our overall evaluation.

Models have been extensively used in simulations of observed climate change during the 20th century. Since the climate forcing, particularly the aerosol forcing, is not perfectly known over that period (Chapter 2), such tests cannot be regarded as unambiguous. For example Knutti et al. (2002) show that in a perturbed physics EMIC ensemble, model versions with a range of climate sensitivities are consistent with the observed surface air temperature and ocean heat content records, if the 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.

Simulations of past climate states allow models to be exercised in regimes that are very different to the present. Such tests complement the 'present climate' and 'instrumental period climate' evaluations, since it could be argued that models can be 'tuned' to reproduce recent climate, and 20th Century climate variations are small compared with the anticipated future changes under SRES forcing scenarios. The limitations of palaeoclimate tests are that both the forcing and the actual climate variables are imperfectly known (and usually derived from proxies), climate states may have been so different (e.g., ice sheets at last glacial maximum) that processes determining quantities such as climate sensitivity were very different to today, and 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.

Finally, climate models can be tested through weather forecasting. Climate models are closely related to the models that are used routinely for numerical weather prediction, 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. The utility of such forecasts suggests that the models used capture certain key processes in the atmosphere and ocean correctly, although the links between these processes and long-term climate response have in many cases not been studied. It must also 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 size and 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 use for projections (see Section 8.4).

8.1.2.1 Integrating component-level and system-level evaluation

The two levels of evaluation described above appear to have little in common. A system-level evaluation lends itself to objectivity and numerical ‘performance indices’ (although the choice of index remains subjective), whereas evaluation of specific parametrisation or numerics choices may be harder to quantify. Is it possible to develop a model that scores well on a system level evaluation (reproduces well a number of climate observations) but gives poor climate change simulations because of deficiencies in its formulation? Experience shows that in some cases model results at the system level can be improved by departing from observationally-justified parameter values. In such cases, errors introduced by incorrect parameter values are presumably compensating for other, as yet unidentified errors in the model.

One way to integrate the two levels of evaluation is to base evaluation on an analysis of those processes that are believed to control climate change response (see, e.g., Sections 8.6 and 8.7). Impact of system-level and component-level errors can then be assessed against their likely impacts on the key processes. Again, such analyses are dependent on the assumption there is no key process that has been omitted from all the driving models.

8.1.2.2 Model intercomparisons

The global model intercomparison activities that began in the late 1980s with the FANGIO (Feedback Analysis for GCM Intercomparison and Observations) project (e.g., Cess et al., 1989) and continued with AMIP (the Atmosphere Model Intercomparison Project; Gates, 1992), 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 coupled model 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 historically observed climate change, and simulations of future climate change. It also differed in that multiple simulations were performed by individual models to make it easier to separate climate change signals from “noise” (i.e., unforced variability within the climate system). Perhaps the most important change from earlier efforts was the collection of a more comprehensive set of model output, which was immediately opened up to the scrutiny of hundreds of researchers from outside the modeling groups. This has led to an unprecedented look at model simulations from a variety of perspectives and has already resulted in a rich offering of submitted journal articles.

The enhancement in diagnostic analysis of climate model results by the broader research community represents an important step forward since the TAR. Overall, the vigorous, ongoing intercomparison activities have several beneficial effects, including increased communication among modelling groups, rapid identification and correction of gross modeling errors, the creation of standardized benchmark calculations, and a more complete and systematic record of modelling progress. A downside is that the effort required of modeling groups to run standardized experiments, prepare output for use by others, and provide model documentation to the community at large impinges on the groups' own research agendas. There is recognition that the limits of model intercomparison activities and standardized experiments should not crowd out other creative research, but some disagreement concerning how resources should be apportioned among them.

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 inevitably 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 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. In the current state of knowledge this is sometimes unavoidable. It is therefore common to adjust parameter values (maybe 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 very 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 observable, then agreement with that observation cannot be used to build confidence in that model. On the other hand, it may be considered subjectively that a model which 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 Chapter 10)

Given sufficient computer time the 'tuning' procedure can in principle be automated using various optimisation procedures; however this has only been feasible to date for EMICs (Annan et al., 2003) and low-resolution GCMs (Jones at al., 2001). Ensemble techniques (Annan et al., 2003; Murphy et al., 2004; Stainforth et al., 2005) allow in principle a range of parameter settings to be generated, each giving equally 'good' climate simulations according to some chosen 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. They also have a potential advantage for certain uses that certain 'emergent properties' (e.g., climate sensitivity) can be specified, so that the sensitivity of model response to that property can be studied (e.g., Knutti et al., 2002 – see Chapter 9). This may be impossible using more complex models since often there is no clear relationship between model parameters and the emergent properties. However a caveat is that key processes, present in more comprehensive models, are not represented in the simplified models, so that a particular result obtained in this way may have limited relevance to the more comprehensive models or to the real world. Little work has been done to date on establishing such ‘traceability’ between EMICs and GCMs.

[START OF QUESTION 8.1]

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

There is considerable confidence that models are reliable enough to provide useful projections of future climate change, particularly at larger scales. 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.

Climate models are mathematical representations of the climate system, expressed as computer codes and run on powerful computers. Model fundamentals are based on physical laws, such as conservation of mass, energy and momentum, along with a wealth of scientific observations. Similar mathematical models are routinely used in many other fields, e.g., to “fly” new aircraft designs before they are built.

Models are able to skillfully reproduce many 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. Comparison typically covers both average climate and its variability, and includes important climate phenomena such as monsoons and the El Niño Southern Oscillation (ENSO). An unprecedented level of evaluation has taken place over the last decade in the form of organised model ‘intercomparisons’-- systematic comparisons of many models against observations as well as each other. Models show significant, and increasing, skill in representing many climate features, particularly at larger spatial scales, and this increases our confidence in their use for simulating future climates. The range of tests also indicates that model skill is real and does not simply arise from adjusting or 'tuning' models to optimise their representation of climate features.

Some climate models, or closely related variants of these models, are also tested by using them to predict weather and make seasonal forecasts. Models are becoming increasingly skilful in this regard, showing that they can represent the important features of the general circulation across shorter timescales, and important features of interannual variability, such as ENSO. Note that limitations in the models’ ability to predict weather beyond a week or so should not be seen as limiting their ability to predict climate changes (see Question 8.2)

Climate models are also able to reproduce many features of past climates and climate changes. Models have been used to simulate paleoclimates, such as the warmer Holocene of 6000 years ago, or last glacial maximum of 21,000 years ago. Within the limitations of paleo reconstructions, they reproduce important features such as the approximate amount of ice age cooling. Models also simulate many observed aspects of climate change over the instrumental record, such as the global temperature trend over the past century (Figure 1), although uncertainties in the magnitude of the cooling associated with sulphate particles provide significant limitations to this test. They can also reproduce features such as the reduction in the diurnal temperature range, and the small global cooling (and subsequent recovery) associated with the Mt Pinatubo eruption of 1991.

[INSERT QUESTION 8.1, FIGURE 1 HERE]

Nevertheless, models still show significant errors, particularly at the regional scale. This is largely because key small scale processes cannot be represented explicitly, and must be included in models in approximate form, as they interact with the larger scale. This is partly the result of limitations in computing power, but also results from limitations in scientific understanding, and in some cases the availability of observations, of the detailed physics of some small scale processes. Examples of this include small-scale buoyancy-driven circulations in the atmosphere and ocean. Significant uncertainties, in particular, continue to be associated with the representation of clouds. As a result, models continue to display a substantial range of global temperature change to greenhouse gas forcing, and assessment of model skill in the representation of current climate has not, to date, been able to significantly reduce this range. Nevertheless, it is noteworthy that models derived separately by many different scientists, without prior knowledge of their response to greenhouse gases, have been unanimous in their prediction of climate warming under greenhouse gas increases.

Since confidence in 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 and climate change. However, as global models continue to develop, and as resolution continues to improve, there are increasing efforts to evaluate and use them for important smaller scale features, such as changes in extremes. As additional computing power becomes available, climate models will be able to resolve regional climate change features more accurately. Models are also becoming more comprehensive in their treatment of the climate system, with recent inclusion of features such as interactive vegetation, ocean biogeochemistry and ice sheet dynamics in some global coupled models. A hierarchy of models with varying degrees of complexity has also been developed, and has proved useful for various applications, such as the projection of very large numbers of emission scenarios.

In summary, confidence in models comes from their physical basis, and their skill in representing observed climate and past climate changes. Models have proved to be extremely important tools for simulating and understanding climate, and there is considerable confidence that they are able to provide useful information on many aspects of future climate change, particularly at larger scales. Models continue to have significant weaknesses, such as the representation of clouds, and this leads to a degree of uncertainty over the magnitude and timing of predicted climate change. Nevertheless, global climate models have provided, consistently over several decades of model development, a robust and unambiguous picture of climate warming in response to increasing greenhouse gases. The warming predicted by models is of a magnitude consistent with what would be expected independently from simple and fundamental physical arguments, observations and paleoclimate reconstructions.

[END OF QUESTION 8.1]

8.2 Advances in Modelling

Since the TAR, many modeling advances have been introduced. They can be grouped into three categories. First, the dynamical cores (advection, numerics, etc.) have been improved, and the horizontal and vertical resolutions of many models have been increased. Second, more processes have been incorporated into the models. This is especially true for aerosol modelling, land-surface modelling and sea-ice modelling. Third, the parameterizations of physical processes have been improved, and new physical processes have been added to the models. For example, a so-called free surface is now widely used in many oceanic models. As a result of these improvements, most AR4 models no longer use flux adjustments (Manabe and Stouffer, 1988; Sausen et al.,1988) to reduce climate drift. This is discussed further in Section 8.2.7.

Although many improvements have been made in individual climate models, many issues remain, and it is impossible to state that any existing model is fully adequate to make projections of the future climate. This is mainly due to the fact that 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.

Brief details of the formulations of each of the AOGCMs used in this report can be found in Table 8.2.1.

8.2.1 Atmospheric Processes

8.2.1.1 Numerics

Since the TAR, efforts have continued to improve the performance of climate models. In the TAR, more than half of the participating atmospheric components used spectral advection. Spectral advection can create spurious tracer quantities such as “negative water” (ref. Williamson and Rasch, 199x). Since the TAR, semi-Lagrangian advection schemes have been adopted in many atmospheric model components. In AR4, both spectral models and grid-point models are being used. Atmospheric model configurations have been changed at several centres; for example, GFDL changed from a spectral model to a grid-point model, while MRI changed from a grid-point model to a spectral model. There is still no consensus on which model configuration is better for the resolutions used in this assessment.

Due to recent advances in parallel computing and strong demand for increased resolution, high-resolution global atmospheric models have been developed at many centres. For such high-resolution models, grid-point methods are considered by many scientists to be most appropriate. Transformations between grid space and wave space become very expensive at high resolution, especially on parallel computers Also, the spectral methods suffer from the so-called Gibbs phenomena, which results from the truncation of the full spectral series. As the resolution of a spectral model increases, the Gibbs phenomena associated with steep mountains and cloud boundaries become non-negligible.

At the same time, however, there are also problems associated with the use of finite-difference methods based on latitude-longitude grids on the sphere at high resolution. These problems include the treatment of the poles and the lack of uniformity and isotropy of the grid. To overcome these problems, new global grid systems have been developed. These include quasi-uniform spherical "geodesic" grids – tessellations of the sphere that are generated from icosahedra or other Platonic solids (e.g., Heikes and Randall, 1995a; Sato et al., 2005), and also a grid based on the conformal "cubed sphere" (McGregor, 1996).

8.2.1.2 Horizontal and vertical resolution

The horizontal and vertical resolutions of the climate models used in AR4 have been increased, relative to the TAR models, by many centres. For example, HadGEM1 has 8 times as many grid cells as HadCM3 (the number of cells has doubled all three dimensions). At NCAR, a T85 version of the CSM is being used in this report, while a T42 version was used in 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 60 level spectral AGCM, which is being used in time-slice mode (Noda et al., 2005). The projections made with these models are presented in Chapter 10.

Due to the increased horizontal and vertical resolution, a number of observed regional climate features as well as global climate features are better reproduced. For example, a far-reaching effect of the Hawaiian Islands in the Pacific Ocean (Xie et al., 2002) has been well simulated (Sakamoto et al., 2004). This is possible now because the Hawaiian Islands can be represented on the grid in the new high-resolution climate model. In addition to such regional climate features, the global-scale climate is also more realistic. The standard deviations of the 500 hPa height on slow time scales are improved in the high-resolution results. Furthermore, the 20-km AGCM at MRI, run in the time-slice mode, can simulate some aspects of regional climate change such as the changes of the Baui front and intensity and number of typhoons (Noda et al., 2005; Kusunoki et al., 2005).

8.2.1.3 Parametrisations

The climate system includes a variety of physical processes, such as cloud processes, radiational processes and boundary layer processes, which interact with each other on many temporal and spatial scales. Because of the limited resolutions of the models, many of these processes are either not resolved or not fully resolved. Because of this truncation, the effects of unresolved processes on resolved processes are accomplished through the use of physical parameterizations. In the past, many parameterizations have been based on simple statistical theories, which typically neglect the effects of process interactions on the small scales. The differences among various parametrization schemes are an important reason why climate model results are different from each other. Although parametrizations have major impacts on the climate model results, these impacts are model-dependent. As an example, a new boundary layer parameterization (Lock et al., 2000; Lock, 2001) had a strong positive impact on the simulation of the GFDL and Hadley Centre climate models, but the same parameterization had less positive impact when implemented in an earlier version of the Hadley Centre model (Martin et al., 2005). This illustrates that 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, 1975, 2004). In recent climate models, microphysical parametrizations are used to predict the distributions of liquid and ice clouds in the atmosphere. These parametrizations can have large impact on the climate sensitivity (Somerville et al.

1990?, Ogura et al., 2005). Realistic parameterizations of cloud processes are considered to be essential to produce good climate simulations and reliable projections of future climate change (see 8.6).

Cloud parameterizations are not simply curve fits or collections of "adjustable parameters". They are physically based 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. Field experiments such as GATE (1974), MONEX (1979), and TOGA-COARE (1993) have been conducted in order to test and improve cloud parameterization schemes. Systematic research such as that conducted by the GEWEX (Global Energy and Water Experiment) Cloud Systems Study (GCSS; Randall et al., 2003 a); 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 AR4 models. For example, the boundary-layer cloud parameterization of Lock et al. (2000) and Lock (2001), mentioned earlier, was tested through GCSS.

Recently, experiments have been performed in which the conventional parameterizations have been replaced with embedded high-resolution models, capable of representing individual large clouds (Grabowski and Smolarkiewicz, 1999; Khairoutdinov and Randall, 2001). The embedded high-resolution models include many more small-scale process interactions. This idea has been variously referred to as the "cloud-resolving convection parameterization" (Grabowski and Smolarkiewicz, 1999) or "superparameterization" (Randall et al., 2003 b). It is hoped that these studies will accelerate the improvement of cloud parameterizations over the coming years. At the same time, an effort has been continued to create large-domain or even global cloud-resolving models. MRI/JMA has run a model with 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). The model makes detailed projections of the evolution of small scale features: in particular, under a 21st Century forcing scenario (SRES A1B) the model suggests that the Baiu front tends to stay around the latitudes of 30–32°N in the western part of Japan, and that years with no end of the Baiu front season often occur. Recently, Sato et al. (2005) reported encouraging results from a global cloud-resolving model. Because of limitations of computer power, it will not be possible to apply global cloud-resolving models to full climate simulations for several more decades.

Aerosols play an important role in the climate system. In some models, sulphate aerosols are specified (e.g., the CNRM model). Fully interactive aerosol models are now used in some models (GFDL _CM2, HADGEM1, CCSR/NIES/FRCGC). In some of these tests, the direct and indirect aerosol effects have been incorporated. 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 discussion is given in Section 8.2.5.

8.2.2 Ocean Processes

8.2.2.1 Numerics

Two of the models used in this report (GISS-EH and BCCR-BCM2.0) include an isopycnic or hybrid ocean vertical coordinate, rather than the more commonly used depth coordinate. Further experience in the use of such models has been gained in recent years, and isopycnic coordinate models can produce solutions for complex regional flows that are as realistic as those obtained with the more common z-coordinate (e.g., Drange et al., 2005). Issues remain over the proper treatment of thermobaricity, which means that in some isopycnic coordinate models the relative densities of, say, Mediterranean and Antarctic Bottom Water masses are distorted. The consequences of such distortion on projections of ocean circulation changes are as yet unknown. The use of such models will contribute to further understanding and narrowing of climate projection uncertainties arising from ocean models. Overall, no clear difference in climate projections has yet been proven to result from the choice of vertical coordinate. Otterå et al. (2004) note that the BCCR-BCM model, which uses an isopycnic coordinate ocean, appears to be considerably less sensitive to high latitude fresh water perturbations that other models with geopotential coordinate oceans. However this may be linked as much to the deep ocean vertical mixing parametrisations (8.2.2.3 below) as to the vertical coordinate type.

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" (unphysical) salt flux. The virtual salt flux method induces a systematic error in sea surface salinity prediction and causes a serious problem at large river mouths (Hasumi, 2002a,b; Griffies, 2004).

Generalized curvilinear horizontal coordinates with bipolar or tripolar grids (Murray, 1996) have become widely used in global ocean models. 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 models have an 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 coupled models.

8.2.2.2 Horizontal and vertical resolution

There has been a general increase in resolution since the TAR, with most ocean models using a horizontal resolution of order 1–2 degrees. Several models use enhanced meridional resolution in the tropics, to better resolve the equatorial waveguide. Eddy-permitting resolution has not been used in a full suite of climate scenario integrations, but since the TAR it has been used in some idealised 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.

Global ocean modeling with resolution high enough to represent mesoscale eddies (e.g., Maltrud and McClean, 2005) has become achievable due to recent progress in computer power. Such models have the advantage of more realistically representing the behaviour of narrow, swift currents, eddy-induced heat and tracer transport, and oceanic short-term variability. A few coupled climate models with eddy-permitting ocean resolution (1/6 to 1/3 degree) have been developed (Roberts et al., 2004; Emori and Hasumi, 2004), and large-scale climatic features induced by local air-sea coupling have successfully been simulated (e.g., Sakamoto et al., 2004). These models have not been used for a comprehensive suite of climate simulations because of the computational cost, but some control and idealized anthropogenic climate change simulations have been made.

Roberts et al. (2004) found that increasing the ocean resolution of the HadCM3 model to 0.33° by 0.33° by L40 (while leaving the atmospheric component unchanged) resulted in many improvements in the simulation of features of the ocean circulation such as those listed above. 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. It was proposed that the full effects of increased ocean resolution would only be seen when the atmospheric resolution was also increased to allow full interaction with the fine scale SST structures.

The Atlantic thermohaline circulation (THC) 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 THC, the fresh Pacific water coming through the Bering Strait is important, and its effect could be wrongly represented without an adequate treatment for its pathway through the Canadian Archipelago and the Labrador Sea (Komuro and Hasumi, 2004). These aspects are improved since the TAR in many of the AR4 models.

Changes around marginal seas are very important for regional climate change. Over these areas, climate is influenced by 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 has a large impact on the regional climate, and the MIROC3.2(hires) model suggests that the Kuroshio axis will be unchanged although its speed is increased as the radiative forcing increases in the future (Sakamoto et al., 2005; see Figure 8.2.1).

[INSERT FIGURE 8.2.1 HERE]

Guilyardi et al. (2004) suggest that ocean resolution may play only a secondary role in setting the time scale of model El Niño variability, with the dominant timescales being set by the atmospheric model provided the basic speeds of the equatorial ocean wave modes are adequately represented (as they typically are with horizontal resolution of order 1 degree)

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 thermoclime structure and meridional overturning (Böning et al., 1995, see 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), KPP (Large et al., 1994), and bulk mixed layer models (e.g., Kraus and Turner, 1967). Representation of the sea surface mixed layer has been much improved due to developments in these parameterizations. 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 THC (e.g., Marotzke, 1997; Hasumi and Suginohara, 1999; Otterå et al., 2004; Oliver et al., 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 must flow over ocean ridges or down continental slopes before connecting to the global THC. 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 processes have come into use in global ocean models (e.g., Nakano and Suginohara, 2002; Winton et al. 1998) and also in some coupled climate models. 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

8.2.3.1 Surface processes

The major advance since the TAR has been the development of terrestrial carbon components in land surface models. Several climate models have evolved to account for the dynamics of the carbon cycle including dynamic vegetation and soil carbon cycling (see Figure 1, TAR Technical Summary). This has led to the incorporation of the terrestrial biospheric feedback in some coupled climate models used in Chapter 10. These feedbacks include the responses of the terrestrial biosphere to increasing $CO_{2,}$ climate change and changes in climate variability (see Chapter 7). Some progress has been made on the specific roles of individual terrestrial feedbacks since the TAR. For example Betts et al. (2004) showed that CO_2-induced reductions in stomatal conductance over the Amazon contributed around 20% of the reduction in rainfall simulated over this region under increasing CO_2 that leads, in part, to Amazonia die back (Cox et al., 2004) which would be a positive feedback enhancing further warming via CO_2 release.

Specific studies have explored the significance of these feedbacks as reported in the TAR (Chapter 7.4.1). Some additional studies have suggested that the biospheric feedback may be significant regionally (e.g. Eastman et al., 2001; Lu et al., 2001; Narisma et al., 2003). A significant change since the TAR is that there has been a series of model evaluation and model sensitivity studies that explore the present modelling capacity of the response of the terrestrial biosphere rather than the response of just one or two components of the biosphere. 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 (see Section 8.3). There is also evidence emerging that regional-scale projection of warming requires the simulation of processes that operate at finer resolutions than current climate models resolve (Pan et al., 2004).

The addition of the terrestrial biosphere models that simulate changes in terrestrial carbon sources and sinks into fully-coupled climate models is still at the cutting edge of climate science. 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) the magnitude of which is currently uncertain (Cox et al., 2000; Friedlingstein et al., 2001; Dufresne et al., 2002) and may be partially related to the climate sensitivity but also to the response of vegetation and soil carbon to increasing CO_2 (Friedlingstein et al., 2003). Thompson et al. (submitted) suggest that the rate at which CO_2-fertilization saturates in terrestrial systems dominates the present uncertainty in the role of biospheric feedbacks and precludes determination of whether the land will act as a negative or positive feedback on increasing CO_2. Joos et al. (2001) and Govindasamy et al. (submitted) find that the response of the biosphere to increasing CO_2 within a climate model depends on the assumed climate sensitivity with high climate sensitivities leading to the biosphere acting to amplify initial warming via a loss of soil and vegetation carbon while at lower climate sensitivities the terrestrial system maintains a dampening effect on warming via CO_2 uptake. Overall, the roles of the terrestrial system and terrestrial biospheric feedbacks remain uncertain but a substantial clarification of the issues has occurred since the TAR. It is now recognised that the terrestrial biosphere may contribute a strong feedback on future climate, and the sign of that feedback depends on the rate and amount of CO_2 increase and the climate sensitivity. The major uncertainties in how to parameterize vegetation and soil carbon processes remain (see Chapter 7).

Most other individual components of land surface processes have been improved since the TAR, such as a sub-grid scale snow parameterization (Liston, 2004), root parameterization (Arora and Boer, 2003; Kleidon, 2004), the representation of high latitude organic soils (Hall et al., submitted) and higher resolution river routing (Ducharne et al., 2003). A recent advance is the coupling of ground water models into land surface schemes by Liang et al. (2003), Maxwell and Miller (2005) and Yeh and Eltahir (2005). These have only been evaluated locally but show promise and may be adaptable to global-scales. In general, the improvements in land surface models since the TAR are based on detailed comparisons of the land surface component against observational data.

Two basin-scale intercomparisons have been performed since the TAR leading to a focus on the modelling of runoff and snow. Boone et al. (2004) used the Rhone Basin to investigate how 15 land-surface models simulate the water balance for several annual cycles compared to data from a dense observation network. They found that most of the land surface schemes simulate very similar total runoff and evapotranspiration for three annual cycles, but the partitioning between the various components varies greatly, resulting in different soil water equilibrium states and simulated discharge. Boone et al. (2004) also showed that more sophisticated snow parameterizations led to superior performance. These provide a general increase in our confidence in the performance of the land surface component of coupled climate models.

A river-routing model (Oki and Sud, 1998) has been incorporated into the Hadley Centre and CCSRNIES/FRCGC models and shown to contribute to the improvement of fresh water flux to the ocean. The use of river runoff as a model evaluation tool has also been demonstrated by Kanae (2005).

An analysis of AMIP-2 results has been performed to explore the land surface contribution to the 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 that capacity depends on improved surface observations, for example, the use of stable isotopes (e.g., McGuffie and Henderson-Sellers, 2004; Henderson-Sellers et al., 2004). Pitman et al. (2004) explored the impact of the level of complexity used to parameterize the surface energy balance on the simulated 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. This suggests that the simulation of mean temperature variance and the variance of extreme temperature are not limited by uncertainties in how to parameterize the surface energy balance. This adds confidence to the use of climate models.

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 dramatically overestimates the likelihood of drought (Seneviratne et al., 2002). However, relatively small improvements to the land surface model, to include variable water holding capacity and a simple canopy conductance for example leads to significant improvements (Milly and Shmakin, 2002). This suggests that most models used in Chapter 10 represent the continental-scale land surface adequately unless warming strongly affects the terrestrial carbon balance (e.g., Cox et al., 2000). Given that some evidence suggests that carbon storage and the physiological and structural responses of the vegetation to increasing CO_2 is extremely important, more coupled climate models with the capacity to capture these processes and a more systematic evaluation of these models, would help increase our confidence in the contribution of the terrestrial surface to future warming.

8.2.3.2 Soil moisture feedbacks in climate models

Both the mean and variability of climate are the result of balances between feedbacks and sensitivities within the climate system. A key role of the land surface is as a store of soil moisture and a control of the evaporation of soil moisture. A potentially important process, the soil moisture-precipitation feedback, has been explored extensively since the TAR building on regionally-specific studies that have demonstrated that links exist between soil moisture and rainfall (e.g., Beljaars et al., 1996; Trenberth and Guillemot, 1996). Recent studies (e.g., Hong and Pan, 2000; Pal and Eltahir, 2001; Georgescu et al., 2003; Gutowski et al., 2004; Pan et al., 2004) support the idea that precipitation simulations during the summer season strongly depend on an surface processes, notably in the simulation of regional extremes (e.g., Schubert et al., 2004a). Douville et al. (2001) showed that soil moisture anomalies affected the African monsoon while Schär et al. (2004) and Black et al. (2004) have both suggested that an active soil-moisture precipitation feedback was linked to the recently anomalously hot 2003 European summer.

The soil moisture-precipitation feedback in climate models had not been systematically assessed at the time of the TAR. It is significantly associated with the strength of coupling between the land and atmosphere (that is, the degree to which temporal variations in land state affect the evolution of weather processes). This coupling strength is not directly measurable at the large scale in nature and has only recently been quantified in land surface models (e.g., Dirmeyer, 2001). Land-atmosphere coupling strength is an important element of the climate system; it is integral, for example, to studies of the climatic impacts of land use change and to the potential for improving seasonal forecast skill through soil moisture initialization.

A recent analysis (Koster et al., 2004) provides a first-order assessment of where the soil moisture-precipitation feedback is regionally important in northern hemisphere summer. They objectively quantified 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 on the globe precipitation in northern hemisphere summer is most strongly affected by soil moisture variations (Figure 8.2.2). These “hot spots” of strong coupling are found, as expected (Koster et al., 2000), in transition regions between humid and dry areas. The models, however, also show strong disagreement in the strength of land-atmosphere coupling. A very small number of studies have begun to explore the reasons for the differences in coupling strength. Seneviratne et al. (2005) highlight the important of differing water-holding capacities amongst the models while Lawrence and Slingo (2005) explore the role of soil moisture variability and suggest that a high occurrence of soil moisture saturation and low soil moisture variability could partially explain the weak coupling strength in the Hadley model (note that “weak” does not imply “wrong” since the real strength of the coupling is unknown).

[INSERT FIGURE 8.2.2 HERE]

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 climate change as simulated by the suite of climate models in IPCC. However, note that at present no assessment has been attempted for seasons other than northern hemisphere summer (in particular, no assessment has been made for the southern 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 local soil moisture measurements on a global scale (Robock et al., 2000) there remain considerable discrepancies between large scale estimates of observed soil moisture (Reichle et al., 2004). This makes evaluating climate models' simulation of soil moisture difficult.

8.2.4 Cryospheric Processes

8.2.4.1 Ice-sheet modelling

Ice sheet models are used in calculations of long-term warming and sea level scenarios, though they have not generally been incorporated in the coupled GCMs used for 21st Century projections in Chapter 10. The models are generally run 'offline', i.e., forced by atmospheric fields derived from high-resolution timeslice experiments, although Huybrechts et al., 2002 and Fichefet et al., 2003 and Ridley et al., () report early efforts at coupling ice sheet models into climate GCMs, and Ridley et al., () 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. See Chapters 4 and 10 for further detail.

8.2.4.2 Sea-ice modeling

Sea-ice components of current AOGCMs usually prognose ice thickness (or volume), area-covered fraction, snow depth, surface and internal temperatures (or energy), and horizontal velocity. Sea ice salinity is still not a prognostic quantity and is treated either as a constant in space and time or defined in terms of a fixed vertical profile.

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.

Sea-ice thermodynamic descriptions of the current AOGCMs have progressed more slowly: normally they include 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. Remarkably, Semtner's (1976) "0-layer model" (one layer of ice with snow parameterized via the albedo) is still being used in some AOGCMs. On the other hand, modelers have begun adopting more sophisticated thermodynamics, 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).

Snow models have advanced significantly, including such physical processes as water and vapor flow, compaction, grain growth, and snow redistribution by wind (Dery and Tremblay, 2004). These advances have not yet been incorporated into AOGCMs, however. 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, is usually included in global models because of its importance in the Antarctic sea ice system. In spite of recent advances (Maksym and Jeffries, 2000), however, snow-ice formation is typically parameterized in only the simplest way (isostatic balance), and the snow-ice is immediately integrated as part of the bulk sea ice in spite of its unique salinity properties.

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 both improves the thermodynamic sea ice simulation and enables a more physical formulation for ice ridging and rafting events, based on energetic principles. Individual categories in a thickness distribution interact with each other in a manner approximating the interaction of individual ice floes, such that thinner ice ridges preferentially. 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 global climate models. Better numerical algorithms used for the ice thickness distribution (Lipscomb, 2001) and ice strength (Hutchings et al., 2004) have been developed for global climate models.

Even with numerous thickness categories, the ice velocity is usually computed for the entire mass in a grid cell, and it all moves with the same speed. This simplification is necessary mainly because of the computational cost of the dynamics models. Advection is itself fairly expensive, and many models rely on first-order upwind schemes, which are diffusive but relatively cheap in terms of computer time. However, the increasing need for additional ice categories or tracers and more accurate advection favor a tendency towards employing second-order advection schemes (e.g., Merryfield and Holloway, 2003; Lipscomb and Hunke, 2004).

Various numerical approaches for solving the ice dynamics equations are being developed. These 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).

Increasingly, functions normally included in the sea ice component of climate models are being moved to the ocean or atmosphere components. This allows tighter coupling of physical processes important for climate feedbacks, such as boundary layer interactions, on shorter time-scales than standard coupling intervals (an hour to a day). It also allows the ice to interact more closely with the ocean in which it floats (Holland, 2003), including a mixing process called keel stirring (Debernard, 2003). Coupling techniques that allow the best ice modeling practices while maintaining close physical interactions between component models remain a challenge (Schmidt et al., 2004).

Progress has been made in stand-alone ice and regional ocean-ice model configurations toward developing more physical parameterizations, such as a dynamic and prognostic salinity profile that includes percolation and flooding; ice aging effects; prognostic ice and snow densities; snow redistribution; melt ponds and associated effect on the radiation balance; melt pond and brine convection; biogeochemistry; interaction of sea ice with ice sheets and icebergs; anisotropic features in the ice such as lead orientation; more physical ridging algorithms; etc. However, it is difficult to rank these developments in importance from the view point of global climate modeling.

8.2.5 Aerosol Modelling and Atmospheric Chemistry

Climate modeling studies using atmospheric aerosols with chemical transport have greatly improved since the TAR. The aerosol global distributions are simulated more precisely through comparisons with accumulated observational data, especially data obtained from satellite sensors (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,GFDL_CM2 and MIROC). Some models also include the indirect aerosol effects (Takemura et al., 2005).

Recently, major advances have been made in non-aerosol chemistry modeling . In the past, most atmospheric chemistry component models used specified winds (such as the Chemical Transport Model (CTM)). Several chemistry models have now been coupled to climate models for process studies. For example, CHASER has been coupled to the CCSR-AGCM (Sudo, 2002), STOCHEM to HadCM3 (Collins et al. 2003) and MOZART to CAM3 (Horowitz et al. 2003). Another important issue is an interaction with aerosol processes. This interaction has been included in CHASER (Sudo et al., 2002) and INCA (Hauglustein et al., 2003) and reasonable results have been obtained. These studies have highlighted feedbacks of climate change on future atmospheric chemistry (See Chapter 7).

However, atmospheric chemistry model components are not included in the climate models used in Chapter10. CCSM3 includes two processes normally found in atmospheric chemistry models, the modification to GHG concentrations by chemical processes, and conversion of SO2 and DMS to sulphate aerosols.

8.2.6 Coupling Advances

A "coupler" couples the various components of a climate model. For example, the OASIS coupler was developed at CERFACS (Terray et al., 1995) and is 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. A distinction is also made between terrestrial, ocean and sea-ice fluxes.

Coupling frequency is an important issue, because fluxes are averaged during coupling interval. Several models use versions of the KPP ocean vertical scheme (Large et al., 1994). This scheme is very sensitive to the wind energy available for mixing. If the models are coupled at a frequency lower than once per 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. Failure to do this could lead to too little mixing energy and hence shallower mixed layer depths. However, high coupling frequency also brings technical issues; in the MIROC model, the coupling interval is 1 hour. In this case, an internal gravity wave is excited in the ocean, and so some smoothing is necessary.

In the AR4, an ensemble technique is often applied for the global simulations of climate change, to assess the importance of initial conditions in climate projections and historical simulations. Due to limited computer resources, typically less than 10 ensemble members are performed for each radiative forcing scenario. To initalize the individual ensemble members, the initial states are taken from random points in a control run with a constant CO_2, or near the end of a previous perturbation integration.

8.2.7 Flux Adjustments and Initialization

Since the TAR, more climate models have been developed which do not use adjustments of the surface fluxes to maintain a stable control climate. As noted by Stouffer and Dixon (1998), the use of 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 models that do not use flux adjustments, the initialization methods tend to be more varied. Many models initialize their oceanic components using data obtained either directly from the Levitus data set (Levitus 1994, 1997, 1998) or from short ocean-only integrations that used the Levitus data set 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 model 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 the preindustrial radiative forcing held constant allowing the coupled system to partially adjust to the preindustrial forcing. The degree of equilibration in the real preindustrial world is not known. Therefore it seems unnecessary to have the preindustrial control fully equilibrated. After the spin-up integration, the start of the preindustrial control is declared and perturbation integrations can begin.

This method produces a relatively consistent set of initial conditions across many models. 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 most climate models now use some variant of the Stouffer et al. method, this situation is now improved.

8.3 Evaluation of Contemporary Climate as Simulated by Coupled Global Models

Due to nonlinearities in the processes governing climate, the climate system responds to perturbations in a way that depends to some extent on its basic state (Spelman and Manabe, 1984). Consequently, in order for models to predict future climatic conditions reliably, they must simulate the current climatic state with some degree of fidelity. How well models must perform in this regard, however, is unknown. In fact, certain aspects of climatic response to external perturbations may be fairly linear and therefore quite insensitive to the basic state. Global mean temperature response to increased greenhouse gases, for example, is roughly proportional to global mean radiative forcing (i.e., the surface temperature responds quasi-linearly to forcing changes), and thus it may not be sensitive to global mean temperature errors of a few degrees Celsius or more.

Nevertheless, poor model skill in simulating present climate indicates that certain physical processes have been misrepresented. The better a model simulates the complex spatial patterns and seasonal and diurnal cycles of present climate, the more likely it is that all the important physical processes have been adequately represented. Thus, when new models are constructed, and almost all the models considered here have been developed since the TAR, considerable effort is devoted to evaluating their ability to simulate today's climate (Meehl et al., 2005). It should be noted, on the other hand, that preliminary studies relying on “perfect model” simulations (e.g., Murphy et al., 2004; Stainforth et al., 2005) show only a weak correspondence between certain measures of model skill and accurate predictions of future climate, so at this time it is impossible to establish minimum threshold criteria that models must meet to be trusted as reliable prediction tools. Increasingly, modeling groups are turning to additional means of evaluating their models by analyzing, for example, various aspects of unforced variability (see 8.4) and individual processes (see 8.2 and 8.6).

In this section, then, the evaluation of models is undertaken not, primarily, to determine which models are qualified to predict future climate change, but to highlight where models generally perform well and to identify their deficiencies. An additional aim is to quantify the evolution in model skill that has been seen over the last several years. Any improvements in model performance can only increase confidence in their predictions.

Much of the assessment of model performance presented here relies on what will be referred to as "CMIP 20th Century simulations," as called for by the ongoing Coupled Model Intercomparison Project (CMIP)[1]. 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-one models considered here (see Table 8.2.1) 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 not all modeling groups chose to archive all of the output fields called for by CMIP.

In the face of the rich variety of climate characteristics that could potentially be evaluated here, focus is derived by considering the elements that most strongly impact the model response to changes in radiative forcing and on those that strongly affect the surface climate. Since the heat and water cycles are directly governed by processes that convey or store energy and water, the multiple facets of model simulated climate will be evaluated in the context of these two fundamental budgets.

8.3.1 Atmospheric Component

8.3.1.1 Surface temperature and the climate system’s energy budget

In order 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 large-scale distribution of annual mean surface temperature is largely determined by the distribution of insolation, which is moderated by clouds and by transport of energy by the atmosphere and to 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.

[1] CMIP is overseen by the WCRP’s Working Group on Coupled Modeling.

8.3.1.1a Surface temperature

Figure 8.3.1a shows with labeled contour lines the observed time mean (1961–1990) surface temperature. The figure represents a composite of surface air temperature over land and sea surface temperature (SST). Also indicated in Figure 8.3.1 is the difference between the mean field simulated by models that performed the CMIP climate of the 20th Century simulation and the observed field. Away from regions where observations are sparse, the absolute difference is, with few exceptions, less than 2 K. Individual models typcially have larger errors, as indicated by Figure 8.3.1b, where for the ensemble of models, the RMS difference between simulated and observed fields is shown. 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 systematic cold bias over land and warm bias over oceans, especially in many sub-tropical and mid-latitude coastal regions. Given that in the annual mean, the temperature range from the coldest to the warmest location on the globe exceeds 50 K, the model errors of order 2 K are relatively small. In fact when spatial scales larger than 250 km are considered, the pattern correlation between the simulated and observed annual mean temperature is typically about 0.98, indicating that models account for a large fraction of the global temperature pattern.

[INSERT FIGURE 8.3.1 HERE]

The error in the multi-model mean pattern is generally smaller than the root-mean-square (RMS) error, calculated over all models, which is shown in Figure 8.3.1b. Still the "typical" model error (i.e., this RMS error) is less than 3 K over most of the globe. Errors over land tend to be larger than the errors over the adjacent oceans, especially in mountainous regions. The largest individual model errors seem related to the location of the sea ice margins in both hemispheres and the simulation of low clouds in the eastern parts of the tropical ocean basins. Both of these problems have the potential to affect the models' response to changes in radiative forcing.

The largest periodically forced climatic pattern of temperature variation is its annual cycle. Figure 8.3.2 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 the extensive land areas of the Northern Hemisphere where the standard deviation generally exceeds 10 K, the models agree with observations within 2 K. The models, as a group, clearly capture the differences between marine and continental environments and also, the increasingly large magnitude of the annual cycle as one moves to higher latitudes, but there is a general tendency to underestimate the annual temperature range over Siberia. This is one example of a general characteristic of current climate models: they are quite accurate in representing the large-scale features of climate, but can be less reliable on the regional and smaller scales.

[INSERT FIGURE 8.3.2 HERE]

As for the diurnal cycle, the difference between daily maximum and minimum surface air temperature is much larger over land (and also better observed) than in the marine environment, so the analysis here focuses only on the continental regions. The diurnal temperature range, zonally and annually averaged over the continents, is generally too small in the models, as illustrated by Figure 8.3.3. Nevertheless the models simulate the local maximum found in the arid, relatively clear subtropical zones. In these regions the shading effect of clouds is reduced, resulting in more rapid daytime warming, and at night there is reduced trapping of surface emissions of longwave radiation, resulting in more rapid cooling. Other effects can be important locally, for example in deserts where dry surface conditions suppress the daytime cooling by evaporation and transpiration. Although Figure 8.3.3 shows that the general character of the diurnal temperature range is well simulated by models, it is not yet known why models generally underestimate its magnitude, although sampling of the diurnal cycle by the radiation algorithms is an issue as is the boundary layer parameterizations used in these models.

[INSERT FIGURE 8.3.3 HERE]

8.3.1.1b Tropospheric and stratospheric temperature

Surface temperature is strongly coupled to 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. More subtly, the vertical temperature structure (along with water vapor and cloud amount) influences the down-welling flux of longwave radiation impinging on the surface, which strongly influences surface temperature because the magnitude of this flux is on average as large as the incident solar radiation. Errors in the atmospheric temperature are of special concern, then, because they indicate model shortcomings. These shortcomings impact both the surface temperature and the model's response to changes in radiative forcing.

Figure 8.3.4 shows cross-sections of the observed zonal-mean, annual mean temperature, and also the difference between the mean of the model results and the observations. The multi-model mean absolute errors are almost everywhere less than 2 K (compared with the observed range of temperatures, when the entire troposphere is considered, spanning more than 100 K). It is notable, however, that near the tropopause at high latitudes, the models are generally biased cold, and this is reflected in the mean model error. This bias is a problem that has persisted for several years, but in general is now less severe than in earlier models. In a few of the models the bias has been eliminated entirely. It is known that the tropopause cold bias is sensitive to several factors, including horizontal and vertical resolution and the treatment of 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.

[INSERT FIGURE 8.3.4 HERE]

8.3.1.1c The balance of radiation at the top of the tmosphere and cloud effects

The primary driver of horizontal 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 temperature impact of the distribution of insolation can be strongly modified by the distribution of clouds and surface characteristics. Each of these factors will now be considered.

If not for clouds, snow cover, and sea ice, climate models should be able to simulate with reasonable accuracy the absorption, scattering and reflection of sunlight (i.e., SW radiation). Figure 8.3.5a shows the annual mean of the zonally averaged observed and simulated outgoing clear-sky shortwave flux at the "top" of the atmosphere (TOA)[2]. For most models the zonal mean errors are less than 10 W m^{-2} except at higher latitudes where models may have difficulties either simulating the distribution of snow and ice or their impact on surface albedo (see Sections 8.3.3 and 8.3.4). Some models do not account for the changes in surface albedo associated with seasonal changes in vegetation, which again primarily affects higher latitudes. The ERBE estimates of clear-sky fluxes are also not as reliable over regions of high surface albedo, so some of the apparent discrepancy between models and observations at high latitudes may be due to observational errors. An additional reason for apparent discrepancies between the model and observations is in the method of obtaining "clear-sky" fluxes. Models typically obtain clear-sky fluxes by executing their radiative code twice, once with the simulated clouds and once with all clouds removed. Observed clear-sky fluxes, on the other hand, are obtained by sampling only cloud-free areas. These different sampling procedures can lead to apparent differences in the clear sky fluxes.

[INSERT FIGURE 8.3.5 HERE]

Clouds increase the total outgoing shortwave radiation by about 50 W m^{-2}, with respect to the annually averaged global mean amount. In the intertropical convergence zone (ITCZ), the total outgoing radiation is more than twice the clear-sky radiation. In the annual mean the Earth in fact appears to be about equally bright at all latitudes, as shown by Figure 8.3.5b. Still the impact of the zonal cloud structure is evident with a local maximum found in the tropics, where the seasonally migrating ITCZ produces relatively high cloud

[2] The atmosphere clearly has no identifiable "top", but the term is used here to refer to an altitude above which the interaction of sunlight with atmospheric molecules becomes trivially small.

amounts, and in mid-latitudes where extra-tropical cyclones and their frontal clouds are formed. At most latitudes, the difference between the mean model zonally averaged outgoing SW and observations is in the annual mean less than 6 W m^{-2}, not much larger than the mean model error in clear-sky flux.

There are, of course, seasonal and east-west variations in cloud cover that are unaccounted for in Figure 8.3.5b, and individual models vary in their ability to simulate these variations. To illustrate this, Figure 8.3.6 shows for each latitude band the model root-mean-square (RMS) error in simulating net TOA shortwave flux, calculated over all longitudes and all 12 months. The errors tend to be substantially larger than the zonal mean errors (cf. Figure 8.3.5b), evidence again that model errors tend to increase as smaller spatial scales and shorter time scales are considered. Figure 8.3.6 also illustrates a common result that the errors in the multi-model average of monthly mean fields are often smaller on average than the errors in the individual model fields. In the case of outgoing SW radiation, this is true at all latitudes. If at each latitude the weighted average of the mean-square error is computed to form a global mean RMS error, the individual model errors are in the range 18–22 W m^{-2}, whereas the error in the multi-model mean climatology is only 13.4 W m^{-2}. Why the multi-model mean field turns out to be closer to the observed than any of the fields comprising it 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* this should be the case. The apparent superiority of the mean model result supports reliance on a diversity of modeling approaches.

[INSERT FIGURE 8.3.6 HERE]

In the annual mean, the net shortwave radiation at the top of the atmosphere is everywhere largely compensated by outgoing LW radiation (i.e., infrared emissions) from the surface and the atmosphere. Globally averaged, this mean annual compensation is nearly exact. The pattern of LW radiation emitted by earth to space depends most critically on surface temperature, atmospheric temperature, humidity and clouds. Figure 8.3.7a shows that the annual mean of the zonally averaged outgoing LW radiation is well simulated by all the models. With a few exceptions the models can simulate the observed zonal means within 10 W m^{-2}. The relatively high humidity and extensive cloud cover in the tropics raises the effective height at which LW radiation emanates to space. Because the temperature is lower at high altitude, the outgoing LW radiation is less than in the subtropics where clearer, dryer conditions prevail.

The seasonal cycle of outgoing LW radiation and the east-west variations of this field are also reasonably well simulated by models. Figure 8.3.7b shows, for each latitude band, the model RMS error in simulating net TOA outgoing longwave radiation (OLR), calculated over all longitudes and all 12 months. The RMS error for individual models varies from about 3% of the OLR near the poles to somewhat less than 10% in the tropics. The errors for the mean of the model simulations are again smaller than the individual models, ranging from about 2% to 6% across all latitudes.

[INSERT FIGURE 8.3.7 HERE]

For a climate in equilibrium, any local annual mean imbalance in the net TOA radiative flux must be balanced by a vertically integrated net horizontal divergence of energy imparted by the ocean and atmosphere. Consequently, the time mean of the zonally averaged poleward transport of energy (by the atmosphere and ocean combined) can be inferred from the net TOA radiative fluxes, and this is shown in Figure 8.3.8. In order for a model to agree well with the observations in this respect (and in the absence of compensating errors) a model must simulate a wide variety of processes correctly. Not only must the atmosphere and ocean transport energy in a realistic manner (which in the atmosphere means that the midlatitude "weather" systems must be correctly simulated), but the models must be able to represent clouds well enough that their impact on TOA radiation is realistic. Although superficially this would seem to provide an important check on models, it is likely that in current models compensating errors do improve their apparent 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). Nevertheless, the degree to which the simulated and observed zonal mean implied energy transports agree is encouraging.

[INSERT FIGURE 8.3.8 HERE]

8.3.1.2 Moisture and precipitation

Unlike temperature, which exhibits large-scale horizontal variations and temporal changes that originate directly from the characteristics of the insolation pattern and the configuration of the continents, precipitation is most directly governed by processes that are internal to the climate system. Although precipitation totals tend to be lower in high latitudes, this is more directly related to temperature than insolation. In addition to the general tendency for warmer air to be moister (due to its higher capacity to hold water vapor), atmospheric transport of water vapor and vertical motion, produced by atmospheric instabilities of various sorts and the flow of air over orographic features, largely determine the distribution of precipitation.

In order for models to accurately simulate the global distribution of the annual cycle of precipitation they should not only simulate evapo-transpiration, but also the many atmospheric processes that move the water vapor around and eventually force it to condense. Many of these processes are difficult to evaluate on a global scale but are discussed further in Sections 8.2 and 8.6. Here the focus will be on the distribution of precipitation and water vapor. The impact of precipitation on the thermohaline circulation is discussed in Section 8.3.2.

8.3.1.2a Precipitation

At the largest scales, annual mean precipitation tends generally to decrease with latitude, reflecting 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. As Figure 8.3.9 shows, however, there is a local minimum in precipitation near the equator, reflecting a tendency for the ITCZ to reside longer in one hemisphere or the other during its annual cycle. 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 because they can adequately account for these features of atmospheric circulation.

[INSERT FIGURE 8.3.9 HERE]

Models also simulate many of the major regional characteristics of the precipitation field. Figure 8.3.10a shows observed annual mean precipitation and Figure 8.3.10b shows the multi-model mean field. The structure of tropical precipitation is similar both in the major convergence zones and also over the tropical rain forests. The signature of extra-tropical cyclones and the effects of warm ocean currents is evident in mid-latitudes, and some topographically induced local precipitation maxima are also simulated by the models (e.g., along the western coastal mountains of Canada). The multi-model mean simulation, however, also is generally deficient in reproducing some of the details of the observed precipitation. 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 broad in most models with too much rain south of the equator.

[INSERT FIGURE 8.3.10 HERE]

Considerable effort has been devoted to examining the tropical Pacific model errors, partly because of the importance of El Niño events that originate there. There is an unrealistic tendency for models to split the intertropical zone into two zones, one north and the other south of the equator. Figure 8.3.11 shows that during the boreal spring, for example, the maximum observed precipitation in a sector of the Eastern Pacific running from 120W to 100W lies north of the equator, with a small secondary maximum south of the equator, but the models reverse the relative sizes of these maxima. Whether or not this common model error affects their projections of global climate change is unknown, but clearly it might not only impact model estimates of changes in precipitation in the tropical regions where it is evident, but also probably influence modes of variability such as ENSO and its influence through teleconnections on mid-latitudes.

[INSERT FIGURE 8.3.11 HERE]

8.3.1.2b Water vapor

Atmospheric humidity is determined by 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 these aspects of the hydrological cycle on global scales is perhaps to determine how well the resulting water vapor distribution agrees with observations. The water vapor distribution is of further interest because it strongly influences the distribution of outgoing LW radiation. Figure 8.3.12 shows the observed annual mean, zonally averaged specific humidity distribution, and also the error in the multi-model mean simulation of that field. A logarithmic scale for labeled contours is used because the the specific humidity decreases roughly exponentially with decreasing pressure. The errors, indicated by the color-filled contours, are expressed as a percentage of the observed value. In the lower tropopsphere the errors in the model mean representation of this field are mostly less than 10%, but nearer the tropopause there is a distinct tendancy for the simulated humidity to be too high in the tropics (by up to 50% for the multi-model mean) and too low at high latitudes (by up to 40% for the multi-model mean). Some of the apparent discrepancy may be due to uncertainty in the observations, especially above the mid-troposphere.

[INSERT FIGURE 8.3.12 HERE]

Any errors in the water vapor distribution should impact the outgoing LW radiation (see Section 8.3.1a.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 strongly affecting the net fluxes at the top of the atmosphere which fundamentally determine climate and climate change. The strength of "water vapor feedback", which strongly affects global climate sensitivity, is, of course, determined by the *changes* in water vapor, 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 storm systems

The cumulative impact of extra-tropical cyclones on particular regions of the extra-tropics derives primarily from their role in transporting heat, momentum and humidity. Extra-tropical cyclones can be both beneficial in providing much of the precipitation for a region and destructive through flooding and damaging winds. Their role in climate change is therefore important.

The first challenge in assessing the ability of climate models to represent extra-tropical cyclones is to determine how to best diagnose and characterize their behavior. A wide range of diagnostic methods have been applied, ranging from simple Eulerian methods (filtered variances) to those based on identifying and tracking cyclones (Blender, 1997; Sinclair, 1994; Murray and Simmonds, 1991; Zolina and Gulev, 2002; Hodges, 1996, 1999). Non-tracking, cyclone counting schemes have also been used (Lambert, 1994; Zhang and Wang, 1997). Cyclone tracking provides the most direct and complete information on extra-tropical cyclones. The results from this approach, however, can depend on the particular method used to identify and track the cyclones, the frequency and resolution of the data, the way statistics are generated from the tracks, and the field used for the identification (Hoskins and Hodges, 2002). The best results are obtained by sampling at high frequency (at least every 6 hours) fields, such as vorticity, that are not dominated by the large scale background (Hoskins and Hodges, 2002; Sinclair, 1994). Past analysis of cyclones in GCM data has been limited by the availability of the high frequency data. Consideration of a variety of variables and levels can also greatly improve our understanding of cyclones in climate models (Hoskins and Hodges, 2002).

The second challenge is validation. The meteorological fields in climate models are generally validated against reanalyses, which are essentially produced using operational systems that assimilate non-homogeneous observations of varying quality. Although for the purposes of mean climate, reanalyses provide the best available observationally-based dataset, there may be problems in data sparse regions. In the Northern Hemisphere, most reanalyses are in close agreement with respect to their cyclone climatologies for the period 1979-present (Hodges et al., 2003; Hanson et al., 2004) and in the case of the ERA40 and NCEP reanalyses even for the longer period 1958-present. In the Southern Hemisphere, however, where observations are dominated by satellites, there is still considerable uncertainty in the representation of cyclones even in the relatively data-rich period (1979-present) (Hodges et al., 2003).

The AMIP II subproject on extra-tropical cyclones (PCMDI, 2004) highlighted the difference in the distribution of cyclones and their properties for a range of climate models. All models diagnosed were capable of producing storm tracks in more or less the correct positions but nearly all showed some deficiency in the distribution and level of activity of cyclones when contrasted with reanalyses. In particular many simulated storm tracks were oriented more zonally than is observed.

In more recent studies, the diagnostic methods of Hoskins and Hodges (2002) have been applied to the higher resolution simulations now available. Examples include analyses of AMIP and coupled model integrations with the ECHAM5 model (Bengtsson et al, 2005) and various versions of the Hadley Centre model (Martin et al., 2004, 2005; Slingo et al., 2002), and, using a different analysis method a study of the Japan Meteorological Agency forecast model, which used a different analysis method (Geng and Sugi, 2003). Slingo et al (2003) emphasize the importance of increasing not only the horizontal resolution, but the vertical resolution as well.

The correct response to changes in SST associated with ENSO is also reproduced well in both AMIP and coupled integrations albeit with a somewhat stronger signal than observed (Bengtsson et al, 2005). As the horizontal resolution is increased in the Hadley Centre atmospheric component model, it shows an increasing ability to represent the sea level pressure signature of extra-tropical cyclones, although it shows some deficiencies when coupled to the ocean (Martin etal., 2005). Lambert and Fyfe (2005) find that, as a group, the coupled GCMs participating in the IPCC AR4 exercise tend to slightly underestimate the number of cyclones in both hemispheres. With regard to intense cyclones, models tend to differ substantially. In general, however, there are a greater number of intense events, both simulated and observed, in the Southern Hemisphere than in the Northern Hemisphere.

Our assessment is that since the last IPCC report, climate models have improved in their ability to correctly simulate extra-tropical cyclone activity and that this is a result of moving to higher resolution and introducing improved model physics. Remaining deficiencies in the representation of these storm systems and their properties appear to be attributable to inadequate specification or simulation of "boundary condition" quantities such as orography, sea ice and SSTs.

8.3.2 Ocean Component Evaluation

Here we focus on fields and variables having a large impact on the magnitude of the climate response when the radiative forcing changes. The magnitude of the surface temperature response is determined by a model’s climate sensitivity and its oceanic heat uptake (Hansen et al., 1984; Rapier et al., 2002). Climate sensitivity in turn is largely determined by atmospheric feedbacks that at any point in a transient integration are related to the magnitude of the sea surface temperature change. The oceanic heat uptake is a function of the vertical heat exchange between the surface and deeper layers in the ocean. The sea surface salinity (SSS), 3-D temperature and salinity distribution and merdional overturning ocean circulation fields are instrumental in determining the magnitude of the oceanic heat uptake and in affecting how the ocean transports heat from one part of the globe to another.

Our analysis involves data obtained from the CMIP 20th Century integrations forced with the modelling groups' best estimates of the historical radiative forcing changes. The model data is compared to observations, mainly taken during the latter part of the 20th Century, although for some fields (SST for example), the observations extend back into the 19th Century. An assessment of the modes of natural, internally generated variability is found in the following subsection (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 issue in the observations. It is our assessment that most of the biases and errors identified below are due to problems in the models’ simulation and not due to these other uncertainties.

Due to space limitations, only a small subset of the analysis to which ocean component models are subjected is discussed here. The discussion below is therefore incomplete. We present mainly summaries and an overview of the analysis. For the individual model analysis and more specialized analysis, please see the supplementary material available on-line.[3]

8.3.2.1 Simulation of mean temperature and salinity structure

Before discussing the oceanic variables, it is important to discuss the fluxes the ocean receives from the atmosphere. In a sense, this is the bridge between the ocean and the atmosphere, which is discussed in the preceding sub-section (8.3.1). These fluxes in large part control the quality of the oceanic simulation. Without reasonably simulated fluxes coming from the atmosphere, the oceanic component will suffer. Of course, this is a coupled problem where the fidelity of the oceanic simulation feeds back on the atmospheric simulation, affecting the surface fluxes.

The total heat flux into the oceans, zonally averaged over all basins, is shown in Figure 8.3.13. An observational estimate is also given, but the net surface heat flux itself is not ordinarily measured; it is inferred from observations of other fields, such as surface temperature and winds. Consequently, the uncertainty in the observational estimate shown in Figure 8.3.13 is large – of the order of tens of W m^{-2}, even in the zonal mean. Within this considerable uncertainty, models appear to be largely consistent with the observations. Both in models and in the observations the total heat flux is a maximum near the equator (about 60 W m^{-2}), with minima found in the subtropics and polar regions. In the Arctic and in the Southern Hemisphere where observations are sparse, the model and observational differences are largest. Also the spread among the model results is larger in these regions.

[INSERT FIGURE 8.3.13 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 the seasonal cycle of surface temperature and shift its phase. The AR4 models evaluated here agree well with the observations of seasonal heat storage by the oceans (Gleckler et al., 2005; see supplemental material). The most notable problem area for the models is in the tropics, where many models continue to have biases in representing the tropical convergence zones. These zones are important pathways where the ocean transports the excess heat it receives near the equator to higher latitudes.

North of 45N, most models transport too much heat northward when compared to observational estimates (Figure 8.3.14). From 45N to the equator, most model northward heat transport estimates lie between the observational estimates. 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 8.3.14 HERE]

The net freshwater flux into the ocean is shown in Figure 8.3.15. Near the equator, the flux is positive due to the Intertropical Convergence Zone (ITCZ). In the subtropics, evaporation is larger than the precipitation fluxes so that the freshwater fluxes are negative. In middle latitudes, evaporation is reduced relative to the subtropics and precipitation increases due to the presence of the middle storm track, so that the net freshwater flux is positive. The scatter among the models is largest in the Northern Hemipshere, at least in part due to the smaller ocean areas in the zonal average.

[INSERT FIGURE 8.3.15 HERE]

As the oceans transport heat, they also transport water (Figure 8.3.16). Near 10°N, the models move fresh water away from the region of maximum fresh water flux. The peak southward freshwater transport occurs around 45°N. The peak northward freshwater transport is found near 40°S. In both hemispheres, the models transport fresh water towards the subtropics (or salty waters away from the subtropics).The spread among the models is largest just south of the equator where many models simulate an ITCZ that is too intense. This is a common model error (8.3.1).

[3] Supplementary material is available to reviewers at the same web site as used for the chapter drafts.

[INSERT FIGURE 8.3.16 HERE]

The annual mean zonal surface wind stress, zonally averaged over the oceans, is reasonably well simulated by the models, as shown in Figure 8.3.17. At most latitudes, the observational estimates 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 observations. In middle to high latitudes, the model simulated wind stress maximum lies equatorward of the observations. This error is particularly large in the Southern Hemisphere. Almost all models place the Southern Hemisphere wind stress maximum north of the observational estimate with the possible exceptions of the CM2.1 and MIROC3.2 (highres) models.

[INSERT FIGURE 8.3.17 HERE]

The Southern Ocean wind stress error has a particularly large negative 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, supplementary materials[3]) is also placed too far north in most models (Russell et al., 2005). Since the Antarctic Intermediate Water (AAIW) is formed on the north side of the ACC, the water mass properties of the AAIW are distorted (typically too warm and salty) by this error. This error contributes to the model mean error identified below where the thermocline is too diffuse, because the waters near the base of thermocline are too warm and salty.

The successes and problems in the surface fluxes noted above contribute to the quality of the simulated oceanic fields discussed below. The largest individual model errors in the zonally averaged sea surface temperature (SST) plots (Figure 8.3.18) are found in middle and high latitudes. One of the largest model mean SST errors is found in middle latitudes of the Northern Hemisphere where the model temperatures are too cold; almost every model in the database has some tendency for this cold bias (see supplementary material[3]). In the model mean, zonally averaged surface temperature error curve, one see some evidence of a warm bias just south of the equator. This may be related to the so-called double ITCZ problem identified in 8.3.1. In the zonal averages near 60S, there is a warm bias in the model mean results. Many models suffer from a too warm bias in the Southern Ocean surface temperature distribution. A similar warm bias exists near the sea ice edge in the Northern Hemisphere (70N), although it should be noted that the areal extent of this latter problem is limited due to the small ocean area found at this latitude.

[INSERT FIGURE 8.3.18 HERE]

In the individual model SST error maps, one also notes 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 model mean result (Figure 8.3.19). This error is associated with problems with the local wind stress, oceanic upwelling and under prediction of the low cloud amounts. The under prediction of the low cloud amounts in these regions may impact the model's climate sensitivity (see 8.6 for more discussion on this point). In spite of these errors, the model simulation of the SST field is fairly realistic overall. Over most latitudes, the model mean, zonally averaged SST error is less than 2^{o}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^{o}K over most regions, with only relatively small areas exceeding this value.

[INSERT FIGURE 8.3.19 HERE]

Compared to SST, sea surface salinity (SSS) is typically much more difficult to simulate. The atmosphere tends to damp SST anomalies, whereas it is essentially unresponsive to SSS anomalies. Furthermore, rivers are an important source of freshwater for the ocean surface. Most of the models use highly idealised river outflow schemes in which the outflow is typically spread over a wider region of adjacent ocean than is seen in reality. (See 8.2 for more discussion on this topic). Because of the small damping of SSS anomalies, any SSS errors that develop may continue to grow until they are quite large, in magnitude and extent.

The zonally averaged, model mean SSS error is less than 0.5 PSU south of 30N (Figure 8.3.20). The model spread over these latitudes is also relatively small. In the middle and high latitudes of the Northern Hemisphere, both the model mean SSS error and the model spread are quite large. North of 45N, the fraction of the latitude circle covered by ocean can be quite small so that relatively small-scale biases can lead to large errors in the zonal average. However taking that fact into account, it is clear that some of the model simulations in these regions are relatively poor.

[INSERT FIGURE 8.3.20 HERE]

Figure 8.3.21 shows the model mean errors in SSS, which are relatively large (1–2 PSU) in the Arctic Ocean where agreement among the models is poor, but most models have a salty bias. The N Pacific Ocean SSS in the model mean field tends to also be slightly saltier than observed. The Atlantic and middle latitudes of the S Pacific and Indian Oceans are generally too fresh in the mean model. There is a wide variety of error patterns among the models. The reader is encouraged to view the supplementary materials[3] for more information. In the Southern Ocean, the SSS is slightly too salty, another error common among many models.

[INSERT FIGURE 8.3.21 HERE]

Problems with the river flow may also be evident in the model mean SSS error map. In a tiny area surrounding the mouth of the Amazon River, the mean model is too salty when compared to observations, perhaps reflecting the fact that some models spread the freshwater input from rivers over an unrealistically large area. There also appears to be a plume of water heading from the Amazon mouth towards the north-northwest where the models are too salty. This error might be explained in part by the tendency of models to rain too little over the Amazon basin (8.3.1). Near the mouth of the Congo River, the model mean SSS error is too fresh. Again, this seems in part related to the precipitation simulation errors in central Africa (8.3.1).

Over most latitudes, the model mean, zonally averaged ocean temperature is too warm throughout much of the ocean depth extending from 200 to 3000 m (Figure 8.3.22). The maximum warm model mean error is located in the region of the North Atlantic Deep Water (NADW) formation in most of the models. The error is about 2^oK. The mean model is too cold above 200 m with maximum cold bias (about 1 C) near the surface in mid-latitudes of Northern Hemisphere. Most models generally have an error pattern similar to the multi-model mean with the exception of CNRM-CM3 and MRI-CGCM2.3.2 which are too cold throughout most of the middle and low latitude ocean. The GISS-EH model is much too cold throughout the subtropical thermocline and only the Northern Hemisphere part of the FGOALS error pattern is similar to the model mean error described here. Please see the supplementary material[3] for individual model and basin averaged error plots.

[INSERT FIGURE 8.3.22 HERE]

The error pattern where the mean model is too warm from about 200 to 3000m in zonal average north of 60S and too cold above 200m, indicates that the thermocline is too diffuse in the mean model. 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 to errors in formation and mixing of North Atlantic Deep Water (see 8.2).

The zonally averaged, ocean salinity plot (Figure 8.2.23) shows that the mean model is too salty in the region of the Mediterranean Sea. In middle and low latitudes, the mean model is too salty compared to observations (Levitus et al., 2005) below 300 m or so. Above 300 m, on the other hand, the mean model is too fresh at many latitudes. The maximum error is about 0.5 PSU at the surface near 15S and by almost 1 PSU near 65N. Individual models typically have even larger problems in these regions of general model bias (Figure 8.3.23b). GISS EH has a very different error pattern when compared to the model mean error with a low latitude fresh bias throughout the ocean and middle latitude salty bias in both hemispheres. The CGCM3.1 does not have the large fresh bias near the surface in the Southern Hemisphere low latitudes. The model mean errors in temperature (too warm) and salinity (too salty) 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 AAIW, as discussed above.

[INSERT FIGURE 8.3.23 HERE]

8.3.2.2 Simulation of circulation features important for climate response

8.3.2.2a Meridional overturning circulation

The meridional overturning circulation (MOC) is an important component of present day climate. This circulation 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 very dense so that it sinks to depth. These waters then flow southward towards the Southern Ocean where they mix with the rest of the World Ocean waters. As the climate warms in experiments where the radiative forcing is increasing, the MOC weakens in many models (Cubasch et al., 2001, chapter 10). The MOC weakening tends to result in a decrease in the northward heat oceanic heat transport in most models (Gregory et al., 2005). The climate changes are also associated with an increase in the vertical stability of the ocean.

The model mean distribution and the simulation obtained from many individual models show a clearly defined overturning circulation connecting the hemispheres where warm, salty surface waters flow into high latitudes of the N Atlantic and return at depth (Figure 8.3.24). The model mean distribution also shows a number of distinct wind driven surface cells. North of 50S, these cells are very shallow. In the latitude of the Drake Passage (55S), the wind-driven cell extends to much greater depth (2 to 3 km).

[INSERT FIGURE 8.3.24 HERE]

Almost all models have some manifestation of the wind driven cells (INM, FGOALS are notable exceptions). The strength and pattern of the overturning circulation varies greatly from model to model. GISS-AOM exhibits the strongest overturning circulation, with almost 40 to 50 Sv. The CGCM (T47 and T63), FGOALS have the weakest overturning circulations, about 10 Sv. The observed value is of order 18 Sv (Ganachaud and Wunsch 2000). Again, the reader is referred to the supplementary material[3] for more details and plots obtained from the individual models.

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.x shows an index of the Atlantic MOC at 30°N, and Figure 10.y the ocean northward heat transport, for the suite of GCM 20th Century simulations. While the majority of models show an MOC strength, and many a heat transport value, that is within observational uncertainty, some show higher and lower values and a few show substantial drifts which would make interpretation of MOC projections using those models very difficult. However no clear relationship has been established between simulated mean MOC strength and the size of the MOC response (e.g., Gregory et al., 2005; Sun, 2005)

Overall, the simulation of the MOC has improved since the TAR, due in part to improvements in mixing schemes and through the use of higher resolution in the oceanic component of the AR4 models (8.2). This improvement is seen in the individual model MOC sections (supplementary material[3]) by the fact that (1) the location of the deep water formation is more realistic, occurring in the GIN and Labrador Seas as evidenced by the larger streamfunction values north of the sill located at 60N (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 N 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 new AR4 models.

8.3.2.2b Southern ocean circulation

Many of the world’s water masses are formed and mix in the Southern Ocean. It is the only basin that spans all the longitudes in a region of prevailing westerly winds. It is also the region where most of the oceanic heat uptake occurs in model integrations subjected to increases in radiative forcing (Sarmiento et al., 1998). Therefore the evaluation of the AR4 model simulations in this region is particularly important.

In most models, the latitude of the Southern Hemisphere zonal wind stress maximum is biased towards the equator (see discussion above). The oceanic current associated with the wind stress maximum, the Antarctic Circumpolar Current (ACC), is also found too far equatorward, causing biases in the simulation of deep and intermediate water formation in the models. The intermediate water subduction zones are located on the equatorward side of the ACC. Shifting these regions equatorward, typically causes the subducted waters to be too warm and salty when compared with the observations (Russell et al., 2005; Kamenkovich and Sloyan, 2005).

It is likely that these errors will influence the transient climate response to increasing greenhouse gases. The Southern Ocean biases in most AR4 models could impact 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. 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 AR4 models. It is notable that this improvement has continued in spite of the fact that nearly all models no longer use flux adjustments (Manabe and Stouffer, 1988; Sausen et al., 1988). This suggests that the improvements in the physical parameterizations and increased resolution noted in 8.2 are having a positive result on the simulation in these models. 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. In the Southern Ocean, the equatorward bias of the westerly wind stress maximum is a problem in most models and this may affect the model's response to increasing radiative forcing.

8.3.3 Sea Ice

The control climate sea-ice conditions are important for determining the magnitude and spatial distribution of the high-latitude warming. Two factors hamper quantitative evaluation of the sea-ice components of the AOGCMs: (1) insufficiency of observations for some key variables (e.g., ice thickness) (see Chapter 4) and (2) pronounced dependence on the errors in simulations of the ice-driving atmospheric and oceanic fields in high latitudes (see Sections 8.3.1, 8.3.2, 11.3.8).

Since the TAR, a major improvement of AOGCM sea ice components in general has been including more sophisticated dynamic components (see Section 8.2, Table 8.2.1). Furthermore, several AOGCMs now include sea-ice thickness categories and relatively advanced thermodynamics. While a dramatic improvement is not obvious in simulations of the current sea-ice climate by AOGCMs as a class (compare Figure 8.3.25 with TAR Figure 8.10; or Kattsov and Källén, 2005, Figure 4.11), some models are now able to better capture key features of sea-ice characteristics geographical distributions and seasonality.

[INSERT FIGURE 8.3.25 HERE]

Sea ice extent (defined as the area poleward of the ice edge) is the most reliably observed sea-ice characteristic (see Chapter 4), and is a primary one for model evaluation. In many models, the sea ice simulated for the current climate is not in agreement with observed coverage, especially when coverage in specific regions is considered (Arzel et al., 2005). Additionally, models show a considerable range even in sea-ice extents for the current climate (Table 8.3.1). At the same time, the multi-model average ice extents are in a reasonable agreement with observations (Arzel et al., 2005; Holland and Raphael, 2005; Zhang and Walsh, 2005). In the NH, the "mean" model slightly overestimates both by the beginning and at the end of the melt season. In the SH, the observed amplitude of the seasonal cycle is too large: the multi-model average ice extent is larger in its seasonal maximum and less in the minimum compared to the observationally based estimates. It appears that the agreement between the models is better in winter than in summer for the both hemispheres, and is generally better in the NH than in SH (Arzel et al., 2005). The biases in the current climate may influence sensitivity of AOGCMs to GHG increase (more pronouncedly in the models with low to moderate (<3) polar amplification, Holland and Bitz, 2003) and to a certain extent confound interpretations of the model-projected coverage for the future time (see Section 8.6).

The spatial distribution of ice thickness varies considerably from one model to another (Arzel et al., 2005). In the absence of reliable climatology for this characteristic, the huge inter-model scatter is by itself an indication of problems inherent in the state of the art AOGCM representations of high latitude processes.

Among primary causes of biases in sea-ice simulations, there are biases in simulations of the atmospheric and oceanic circulation in the high latitudes. For example, sea level pressure (SLP) biases over the polar oceans (e.g. Walsh et al., 2002; Chapman and Walsh, 2005) suggest that the wind fields driving the AOGCM sea-ice components are likely to be responsible for a significant part of the biases in simulated geographical distributions of sea-ice mass and velocities (Bitz et al., 2002), however advanced the sea-ice model dynamics might be. So are surface heat fluxes, whose errors may result in particular from inadequate parameterizations of atmospheric boundary layer (under stable conditions such as over ice, in the night, and in the wintertime), generally poor simulation of high latitude cloudiness which demonstrates a striking inter-model scatter (e.g. Kattsov and Källén, 2005), etc.

Table 8.3.1. Coupled (IPCC AR4) model simulations for March and September (1980–1999) of sea ice extent (10^6 km^2). For each model, sea ice is ascribed to be present in a grid cell if its quantity exceeds 15% concentration – the *ad hoc* ice edge. The "observed" values are based on HadISST data (Rayner et al., 2003). In the brackets, the total area covered by sea ice (the sum of ocean grid cell areas multiplied by the corresponding sea ice concentration values) is given.

Model Name	NH March	NH September	SH September	SH March
CCSM3	19.4 (16.7)	8.7 (5.7)	26.2 (20.5)	7.0 (4.7)
CGCM3.1(T47)	17.3 (15.4)	9.9 (7.1)	28.7 (22.0)	9.1 (5.1)
CNRM-CM3	18.6 (16.3)	8.5 (6.5)	21.2 (17.7)	0.6 (0.2)
CSIRO-Mk3.0	17.5 (15.5)	11.7 (9.7)	20.6 (16.4)	5.5 (2.7)
GISS-AOM	14.6 (12.9)	8.7 (5.6)	24.2 (15.4)	4.7 (1.8)
GISS-ER	17.8 (15.9)	13.3 (11.7)	19.2 (12.7)	2.2 (1.3)
INM-CM3.0	16.4 (12.6)	5.7 (3.9)	25.4 (18.2)	5.5 (3.6)
IPSL-CM4	17.3 (14.6)	8.3 (5.9)	21.0 (15.4)	2.1 (0.9)
MIROC3.2 (hires)	14.7 (12.5)	3.8 (2.3)	20.7 (16.9)	3.2 (1.7)
MIROC3.2 (medres)	17.0 (14.9)	9.6 (7.7)	16.6 (12.2)	1.7 (1.2)
MRI-CGCM2.3.2	17.8 (16.0)	9.6 (8.2)	21.8 (18.5)	5.7 (3.7)
PCM	23.8 (19.1)	11.3 (8.1)	31.1 (19.7)	7.4 (5.2)
UKMO-HadCM3	17.6 (15.1)	6.1 (3.9)	20.9 (17.2)	3.9 (2.0)
UKMO-HadGEM1	18.5 (16.4)	8.3 (5.9)	23.5 (18.8)	6.1 (4.2)
Model mean	**17.7 (15.3)**	**8.8 (6.6)**	**22.9 (17.3)**	**4.6 (2.7)**
Observed	**16.6 (14.2)**	**7.5 (5.6)**	**20.0 (16.1)**	**5.6 (3.1)**

8.3.4 Land-Surface Component

Our capacity to evaluate the land surface component in coupled models is severely limited by the availability of observational data. The key roles of the terrestrial surface are the partitioning of available energy between sensible and latent heat fluxes, the partitioning of available water between runoff and evaporation, snow cover and the exchange 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

Simulations of snow cover over Northern Hemisphere lands by a suite of Atmospheric General Circulation Model (AGCM) experiments submitted by an international array of research groups participating in the second phase of the Atmospheric Model Intercomparison Project (AMIP-2), as well as by the coupled climate models included in Chapter 10, have been evaluated. Evaluations of both snow covered area (SCA) (Frei et al., 2003) and snow mass, or water equivalent (SWE) (Frei et al., 2005), indicate that AGCM snow simulations exhibit significant between model variability, and that the median result from a suite of models is usually more realistic than the result from any one particular model. With regards to SCA, at continental to hemispheric scales AMIP-2 models exhibit improvements over AMIP-1 models, including the elimination of temporal and spatial biases in simulations of the seasonal cycles over both North America and Eurasia. However, biases from individual models can be significant. Over Eurasia, regions are identified where models consistently either under- or over-estimate SCA at the southern boundary of the seasonal snow pack. The region of greatest model bias is eastern Asia, where models overestimate snow extent by ~10^6 km^2.

Since the TAR, intermodel consistency in simulating snow cover has, at least in some respects, increased, as illustrated by Figure 8.3.26 (compare with TAR Figure 8.11). For the Northern Hemisphere, Figure 8.3.27

summarizes the areal extent of terrestrial snow cover in February, as simulated by eight models that performed the CMIP 20th Century experiment. In most regions the majority of the models predict snow where it is observed, but there is also a tendancy for a minority of models to exaggerate the snow area, and nearly all models simulate too much snow over eastern Asia. Frei and Gong (submitted) explored the capacity of the coupled models included in Chapter 10 to simulate winter North American SCA. They found that the range of simulated mean values in these coupled experiments approximated the AGCM findings from AMIP-2, although there is a greater tendency to underestimate mean SCA. Most models were able to simulate the observed decadal scale variability over the 20th century, although the variability is unrealistically dampened in ensemble mean results compared to results from individual ensemble members (Figure 8.3.27).

[INSERT FIGURE 8.3.26 HERE]

[INSERT FIGURE 8.3.27 HERE]

With regards to SWE, Frei et al. (2005) found that AMIP-2 AGCMs simulate the seasonal timing and the relative spatial patterns of continental scale SWE over North America fairly well. A tendency to overestimate the rate of ablation during spring was however identified. On the continental scale, the peak monthly SWE integrated over the North American continent in AMIP-2 models varies between ±50% of the observed value of ~1500 km^3. The magnitude of the model errors is large enough to potentially affect continental water balances.

Further analysis of SCA has been provided by Roesch and Roeckner (submitted), who evaluated surface albedo and snow cover in the recent CMIP 20th Century simulations. Focusing first on the seasonal cycle, they found that most models simulate excessive snow mass in spring and suffer from a delayed spring snow melt, whereas the onset of the snow accumulation is generally well captured. At continental scales, the seasonal cycle of SCA is captured reasonably well by most models. Year-to-year variations are often underestimated in Eurasia in winter and spring, while reasonably well simulated over North America. The surface albedo over snow-covered forests is generally too high in these models.

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) in part due 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 (2001a) used the AMIP-2 framework to show that the Canadian Climate Model's simulation of the global hydrological cycle compared well to observations, but regionally variations in rainfall and consequently runoff led to differences in basin-scale quantities. These errors were attributed principally to errors in the precipitation simulations which is supported by simulations of stream flow that, when driven with observed precipitation compare well to observed stream flow (Arora, 2001b). Gerten et al. (2004) evaluated the hydrological performance of the LPJ model and showed that the model compared well in the simulation of runoff and evapotranspiration compared to other global hydrological models although it is noteworthy that 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. (submitted) used results from Chapter 10 models to investigate whether observed 20th-century trends in regional land hydrology could be attributed to variations in atmospheric composition and solar irradiance. An ensemble of 26 integrations from nine climate models was used covering the 20th Century. They showed that these models simulated observed stream flow measurements at regional scales with good qualitative skill. Further, the models demonstrated highly significant quantitative skill in identifying the regional runoff trends indicated by at 165 long-term stream gages. They concluded that the impact of changes in atmospheric composition and solar irradiance on observed stream flow was partially predictable using Chapter 10 climate models. This is an important scientific advance: it suggests that despite many limitations and weaknesses that remain 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.

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 on a global scale (Robock et al., 2000) considerable discrepancies remain between large scale estimates of observed soil moisture (Reichle et al., 2004). This makes evaluating climate models' simulation of soil moisture difficult.

8.3.4.3 Surface fluxes

Solar radiation is the driving force of the Earth's climate, and despite considerable effort since the TAR, uncertainties remain in its representation in climate models (Potter and Cess, 2004). The major systematic evaluation of the capacity of climate models to simulate solar radiation used AMIP-II climate model data (Wild, 2005) which included many climate models included in Chapter 10. Wild (2005) evaluated these models and showed a considerable degree of difference in the global annual mean solar radiation absorbed at the Earth's surface. Figure 8.3.28 shows substantial differences between the climate models in the atmospheric and the surface absorption of solar radiation. In comparison to global surface observations, Wild (2005) concludes that a large number of climate models overestimate surface absorption of solar radiation due in part 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). A result of the difficulties in simulating absorbed solar and infrared radiation at the surface is an inevitable uncertainty in the simulation sensible and latent heat fluxes at the surface.

[INSERT FIGURE 8.3.28 HERE]

8.3.4.4 Carbon

A major advance since the TAR is that there have been some systematic assessments of the capability of land surface models to simulate carbon. Dargaville et al. (2002) evaluated four global vegetation models' capacity 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 capture the net exchange of carbon and then evaluated the carbon fluxes, via an atmospheric transport model, against 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 were systematically better than the others in their capacity to simulate the global carbon budget, but all four were able to reproduce 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 my observed climatologies. An evaluation of IBIS has also been performed, coupled to a climate model. Delire et al. (2003) coupled IBIS to the NCAR CCM3 and compared the resulting climatology to one produced by IBIS forced off-line with an observed climatology. The climate simulated by the NCAR CCM3 included some biases, which strongly affected the prediction of vegetation. These biases were in part systematic weaknesses in the NCAR CCM3 (northern hemisphere winter cold bias), but were also in part associated with vegetation feedbacks (boreal summer cold bias over Alaska and northern Siberia), which may have been related to the absence of lakes, wetlands and crops in IBIS. The simulation of global biomass and soil carbon appeared reasonable. Overall, Delire et al. (2003) conclude that many of the biases they found in the CCM3-IBIS simulations could be attributed to the atmospheric model. These were both enhanced and reduced via coupling to the vegetation model but overall, the changes in the biases that could be attributed to the coupling to IBIS were small in comparison to the size of the bias. Delire et al. (2002) also compared the simulation of IBIS with a land surface scheme that did not include interactive vegetation coupled into NCAR CCM3. A variety of differences were identified, but these were small enough to provide confidence in the capability of IBIS coupled to a climate model.

8.3.5 Tracking Changes in Model Performance

During the development of new versions of coupled models, it has become common practice to consider the individual submodels (e.g., atmosphere, ocean, sea ice) comprising the climate system and to evaluate each of them in turn in an experimental configuration that isolates insofar as possible that component from the others. The Atmospheric Model Intercomparison Project (AMIP) experimental protocol, for example, has been adopted as a standard way to test the atmospheric component model, independent of the ocean model. In AMIP experiments, the monthly mean sea surface temperature and sea ice area fraction are specified as observed over recent decades. Consequently, most model errors seen in AMIP simulations arise from problems with the atmosphere, not the coupled atmosphere-ocean system, so their fundamental causes can usually be identified more easily than errors apparent only in the coupled system.

Starting around 1992, under the coordination of AMIP, modeling groups have archived output from their AMIP simulations at the Program for Climate Model Diagnosis and Intercomparison (PCMDI). This database of model output has been opened up to scientists outside the groups developing these models. By virtue of the relative accessibility of AMIP output, model behavior is being scrutinized from various points of view by an increasing number of researchers with a very broad range of expertise. Building on the success of the AMIP approach, a Coupled Model Intercomparison Project (CMIP) was established in 1996, which encouraged groups to submit results from their coupled models of the atmosphere and ocean. The standard CMIP experiments are: 1) a control simulation with no changes in atmospheric composition and no interannual changes in solar forcing, and 2) a perturbed simulation with atmospheric carbon dioxide concentration increased at a rate of 1% per year until it doubles. In recent years an increasingly comprehensive subset of model output has been collected and analyzed from the benchmark CMIP experiments. The latest CMIP-coordinated effort has led to the unprecedented collection of coupled model output that is the focus of much of this assessment report. In addition to the standard CMIP experiments, an historical run (i.e., the so-called “20th Century simulation”) and several future scenario runs have been made available to hundreds of research scientists.

One consequence of the standardization of benchmark experiments, exemplified by AMIP and CMIP, is that changes in model performance can now be more easily assessed. Although the most important metrics by which progress might be tracked depend to some extent on the intended applications of the models, there is general agreement that a wide variety of variables should be considered and a broad range of phenomena should be analyzed. Relative to CMIP, the more mature state of the AMIP database and the more complete collection of model output available from both the early and more recent AMIP contributions make it easier to assess changes in the skill of uncoupled atmospheric model components than of the fully coupled system. For this reason, we focus on AMIP simulations in the rest of this section, but we note that we can expect similar analyses will soon appear, based on the output of fully coupled models, collected prior to the TAR and again from more recent models.

To summarize the evolution of the collective ability of atmospheric component models to simulate the mean climate state, Figure 8.3.29 displays metrics of model performance in a Taylor diagram (Taylor, 2001). Statistical comparisons between several simulated and observed fields were made to obtain an overall sense of whether models, following the AMIP protocol, had or had not improved over the decade from 1992–2001. The statistics shown on the diagram are the correlation coefficient between the observed and simulated field (related to the azimuthal angle), the root-mean-square (RMS) difference between the two fields (proportional to the distance to the point on the x-axis marked observed), and the standard deviation (SD) of the simulated field (proportional to the radial distance). The dimensional statistics (RMS error and SD) have been normalized by the observed SD, and the RMS error is computed after removal of generally negligible global mean biases. Statistics are shown based on output from the nineteen modeling centers that reported results from both earlier and later versions of their models. The statistics obtained from the collection of older model versions determine the position of the tails of the arrows, and the arrows point to results obtained from the newer model versions. On this kind of diagram, model improvement is indicated by increasing correlation, reduced distance to the point marked "observed," and decreased distance from the dotted arc (which is located at the observed SD).

[INSERT FIGURE 8.3.29 HERE]

The composite multi-model median result was calculated considering monthly mean output from the ensemble of nineteen models. Output from each model was interpolated to a common grid of 64 latitudes by 128 longitudes. For each grid cell and for each of the 120 months of the decade, 1979–1988, the median of the nineteen model values was then selected. The collection of these values defined the composite multi-model median result. It differs from simply taking the mean of all nineteen model results (at each grid cell and for each month) in that outliers have reduced influence.

The statistics shown in Figure 8.3.29 are the so-called space-time statistics for seasonal data, weighted by the area of each grid cell. In the case of the RMS error, for example, the sum of the squared difference includes contributions from all grid cells (weighted by the grid-cell area) and also all 40 seasons, so the fidelity of the full annual cycle of the spatial pattern is measured, along with interannual variability. It should be noted that the statistics calculated for the composite multi-model median fields are not the same as the median (or mean) of the statistics calculated from the individual model output fields. In fact the agreement with observations of the composite multi-model median field is generally better than the agreement of any of the individual fields from which the median was calculated (see, for example, Figure 8.3.20).

The statistics shown in Figure 8.3.29 characterize how model skill has evolved in simulating the eleven global fields listed in the figure caption. The impression given by the diagram is that models have generally been improved during the decade, 1992–2001, but the fractional decreases in RMS error are generally quite small. This conclusion applies to the composite multi-model median result, but further analysis demonstrates that many individual models have also improved. Figure 8.3.30, for example, shows how individual model skill has generally improved in simulating the annual cycle of the global precipitation pattern, in many cases much more than the composite multi-model median result. The figure also shows that the multi-model median field is more similar to the observed than nearly all the individual model fields, reinforcing the emerging generalization that the collection of models can simulate climatology better than individual models.

[INSERT FIGURE 8.3.30 HERE]

Our appraisal of changes in model performance has focused on the atmospheric component model and is limited in that less than a dozen fields were considered and statistics characterizing only the annual cycle of the global pattern were evaluated. Model skill in simulating interannual variability as well as their skill in simulating the climate of specific geographical regions has not been specifically addressed, but the improving statistics for the mean climatology are certainly encouraging. Although a truly comprehensive evaluation of model skill in simulating contemporary climate would include consideration of a much wider range of characteristics than is possible here, it should be noted that the strong interactions among all the elements that comprise climate means that errors arising anywhere in the system will to some extent be reflected in every variable. The physics governing large-scale midlatitude motion in the atmosphere (well-approximated by the hydrostatic and geostrophic balance) implies, for example, that the polar cold bias evident in many models (see Figure 8.3.4) will be accompanied by errors is the vertical structure of zonal wind (not shown). It is therefore perhaps not unreasonable to characterize overall changes in model performance by examining a limited subset of model fields.

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 climate models used for climate predictions of the future.

8.4.1 Northern and Southern Annular Modes (NAM and SAM)

The Northern Annular Mode (NAM, Thompson and Wallace, 1998; also called the Arctic Oscillation) is a hemispheric-scale pattern that represents the leading mode of variability in the Northern Hemisphere extratropical atmospheric circulation. The NAM is not zonally symmetric, with strongest variations evident over the Atlantic sector where it is closely related to the North Atlantic Oscillation (NAO; Hurrell, 1995). There is evidence (e.g., Fyfe et al., 1999; Yamaguchi and Noda, 2005) that the simulated response to greenhouse gas forcing has a pattern that partly resembles the models' leading modes of variability, and thus it would appear important that these modes are realistically simulated. Analyses of individual coupled GCMs (e.g., Fyfe et al., 1999; Furevik et al., 2003; Holland 2003; Liu et al., 2004; Min et al., 2005) have demonstrated that they are capable of simulating many aspects of the NAM and NAO patterns including linkages between circulation, temperature and precipitation. 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 most recently developed models (Miller et al., 2005; Yamaguchi and Noda, 2005) confirm the overall skill of coupled GCMs but also identify that teleconnections between the Atlantic and Pacific Oceans are stronger in many models than is observed (Osborn, 2004), though some models are biased towards a strong polar vortex in all winters and thus their simulations nearly always reflect behaviour that is only observed at times with strong vortices (when a stronger Atlantic–Pacific correlation is observed, Castenheira and Graf, 2003).

Most models underestimate the observed temporal variability of atmospheric pressure, but organize too much of this variability into the NAM and NAO (Miller et al., 2005). The year-to-year variance of the NAM or NAO is well simulated by some coupled GCMs, while others are significantly too variable (Osborn, 2004); for the models that simulate stronger variability, the persistence of anomalous states is also greater than observed (AchutaRao et al., 2004). The magnitude of multi-decadal variability (relative to sub-decadal variability) is lower in coupled GCM control simulations than is observed, though this may indicate there is an influence of external forcings rather than that the models are in error. 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 veracity of model behaviour.

The Southern Annular Mode (SAM, Thompson and Wallace, 1998; also called the Antarctic Oscillation) is a hemispheric-scale pattern that represents the leading mode of variability in the Southern Hemisphere extratropical circulation. Like its Northern Hemisphere counterpart, the NAM, the SAM has signatures in the tropospheric circulation, the stratospheric polar vortex, midlatitude storm tracks, ocean circulation, and sea ice. Coupled GCMs generally simulate the SAM realistically (Fyfe et al. 1999; Kushner et al. 2001; Cai and Watterson 2002; Hall and Visbeck 2002; Marshall et al. 2004; Holland and Bitz 2005; Delworth et al. 2005). For example, Delworth et al. (2005, Figure 1) compares the SAM in the NCEP Reanalysis to the same in the GFDL/CM2.0 and CM2.1 coupled GCM simulations. The main elements of the SAM spatial pattern, including a low-pressure anomaly over Antarctica, high-pressure anomalies equatorward of 60°S, and a surface warm anomaly over the Antarctic peninsula, are captured well in both versions of the GFDL coupled GCM.

Raphael and Holland's (2005) survey of the simulated SAM in several IPCC AR4 models (GFDL CM2.1, CCSM3, CSIRO-Mk3.0, GISS-ER, UKMO-HadCM3 and MIROC3.2) for the period 1960–1999 casts some doubt on this positive assessment of coupled models' ability to accurately simulate the SAM. In particular, Raphael and Holland find that structural details such as the simulated amplitude, zonal structure, and temporal spectra of the SAM in these models do not always compare well with the SAM in the NCEP Reanalysis. But these structural details vary considerably among different realizations of multiple-member ensembles, and the SAM in the NCEP Reanalysis is problematic when compared to station data (Marshall 2003). Thus it is difficult to assess whether Raphael and Holland's analysis points to a significant shortcoming in the ability of coupled models to simulate the SAM, or if the problem is sampling in the observed analysis.

Resolving these issues may require a better understanding of SAM dynamics. The SAM is primarily a tropospheric phenomenon that can be captured, for example, in atmospheric GCMs with a poorly resolved stratosphere and driven by prescribed SSTs (e.g., Limpasuvan and Hartmann, 2000; Cai and Watterson, 2002). Even much simpler atmospheric models with one or two vertical levels produce SAM-like variability (Robinson, 1991; 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, coupled GCM simulations suggest strong SAM-related impacts on ocean temperature, ocean heat transport, and sea-ice distribution (Hall and Visbeck, 2002; Holland and Bitz, 2005); these could easily implicate air-sea interactions in 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

The Pacific Decadal Oscillation (PDO) is the leading mode of decadal variability in the North Pacific. The PDO has a structure in the atmosphere and upper North Pacific Ocean that resembles the pattern normally associated with ENSO's impact on the region (Latif and Barnett, 1996; Mantua et al., 1997; Zhang et al., 1997; Deser et al., 2004). There are two key differences between the PDO and ENSO. First, the PDO has greater variability in mid-latitudes than it does in the tropical Pacific, whereas for ENSO this hierarchy is reversed (Latif and Barnett, 1996; Zhang et al., 1997; Mantua et al., 1997). Second, the PDO has a corresponding time-series that is more heavily influenced by variability at decadal and longer time-scales than are traditional ENSO indices (Mantua et al., 1997; Newman et al., 2003).

Latif and Barnett (1994) argued that the PDO-like mode they examined could be understood in terms of mid-latitude atmosphere-ocean interactions, without the need for teleconnections with the tropical Pacific. However, more recent work indicates that the PDO 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; Linsley et al., 2000; Evans et al., 2001; Mantua and Hare, 2002; Folland et al., 1999, 2001, 2002; Deser et al., 2004). The appearance of the IPO as the leading EOF of SST in coupled GCMs that do not include interdecadal variability in natural or external forcing (e.g., variability in solar insolation or changes in greenhouse gases) indicates that the IPO is an internally generated, natural form of variability. Note, however, that some models exhibit an El Niño-like response to global warming (Cubasch et al., 2001) that can take decades to emerge (Cai and Whetton, 2001). Therefore some, though certainly not all, of the variability we have seen in the IPO and PDO indices might be anthropogenic in origin (Shiogama et al., 2005).

Coupled models do not seem to have difficulty in simulating IPO-like variability (e.g., Meehl and Hu, 2004; Yeh and Kirtman, 2004), even in models that are too coarse to properly resolve equatorially-trapped waves thought 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 coupled model 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. The importance of properly resolving coastally-trapped waves (Gill, 1982) in the context of simulating decadal variability in the Pacific has been raised in a number studies (e.g., Meehl and Hu, 2004; White et al., 2003; Pierce et al., 2000). It is unclear if the deficiencies in IPO-like pattern in the Pierce et al. (2000) model is unique to their model, or if they arise from sampling or observational error. Further, there has been little work evaluating the amplitude of Pacific decadal variability in coupled models. Manabe and Stouffer (1996) showed that the variability has roughly the right magnitude in their model but a more detailed investigation using more recent models with a specific focus on IPO-like variability would be useful.

8.4.3 Pacific-North American (PNA) Pattern

The Pacific-North American (PNA) Pattern (Wallace and Gutzler, 1981) is a recurrent wintertime circulation pattern in the middle and upper troposphere, with quasi stationary centers of action spanning the North Pacific and North American sectors. This wave-like spatial pattern exerts a notable influence on seasonal changes in temperature, precipitation and synoptic-scale activity over the extratropical North Pacific and North America. The occurrence of the PNA pattern has been attributed to both external and internal factors. 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 (Prediction Of climate Variations On Seasonal to interannual Time-scales) and the U.S. DSP (Dynamical Seasonal Prediction) projects. The skill of seasonal hindcasts produced by the participating models of the 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 atmospheric climate models do respond to the prescribed SST forcing. 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.

The performance of the dynamical seasonal forecast system at the U.S. National Centers for Environmental Prediction in predicting the atmospheric anomalies in the PNA sector has been assessed by Kanamitsu et al. (2002). This system uses a 2-tiered approach, which entails prediction of the tropical Pacific SST using a coupled model, and prescription of the anomalous SST forcing thus obtained as boundary conditions for atmospheric GCM integrations. During the outstanding El Niño event of 1997–1998, the operational forecasts based on this system with one-month lead time are in good agreement with the observed geopotential height, surface temperature and precipitation changes in the PNA sector. More recently, hindcast experiments have been launched using coupled GCMs. 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 (i.e., at a 2–4 month lead relative to the verification period), 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 to be 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 for longer lead times (e.g., 4–6 months). 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 coupled GCMs. 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 of such coupled experiments indicates that the ENSO events appearing in the integration are linked to a PNA-like pattern in the upper troposphere. The centers of action of the simulated patterns are systematically displaced 20–30 degrees of longitude westward relative to the observed positions. This discrepancy is evidently linked to a corresponding spatial shift in the ENSO-related SST anomaly center simulated in the tropical Pacific. This finding illustrates that the spatial configuration of the PNA pattern in coupled models is crucially dependent on the accuracy of ENSO simulations in the tropics.

8.4.4 Cold Ocean-Warm Land (COWL) Pattern

The Cold Ocean-Warm Land (COWL) Pattern (Wallace et al., 1995) is obtained by regressing local surface temperature anomalies on time series of Northern Hemisphere mean temperature. This analysis reveals 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 (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; Corti et al., 1999; Wu and Straus, 2004b; Lu et al., 2004).

In their analysis of coupled model simulations, Broccoli et al. (1998) found that the original method for extracting the COWL pattern could yield ambiguous results when applied to a simulation forced by past and future variations in anthropogenic forcing. 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 Northern Annular Mode (NAM) and a mode closely resembling the Pacific-North American (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

Persistent or recurrent structures of atmospheric circulation are often denoted as climate or weather regimes. These structures have been demonstrated to have considerable effects on surface weather (e.g., Plaut and Simonnet, 2001; Trigo et al., 2004; Yiou and Nogaj, 2004). Weather 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. Therefore it is important to evaluate persistent or recurrent structures. A number of different statistical techniques have been used to characterise these regimes, all designed to diagnose non-Gaussian structure in data (e.g., Ghil and Robertson, 2002; Monahan et al., 2003; Crommelin, 2004); Teng et al. (2004) emphasise the potential sensitivity of such structure to time filtering. 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). On a sectorial scale, 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. These studies have provided evidence that regime structures may be slightly changed, but are not fundamentally altered, by imposed forcing (e.g., SST and greenhouse gases); this result is broadly consistent with the ideas of Corti et al. (1999). Since the TAR, agreement between different studies has improved regarding the number and structure of both hemispheric and sectorial atmospheric regimes, although this remains a subject of research (e.g., Wu and Straus, 2004) and the statistical significance of the regimes has been questioned (e.g., Hannachi and O'Neill, 2001; Hsu and Zwiers, 2001; Stephenson et al., 2004).

An important class of sectorial weather regimes are blocking events, associated with local reversals of the midlatitude westerlies. The most recent systematic intercomparison of 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 less frequent than observed events (e.g., Pelly and Hoskins, 2003b). However, no commonly accepted objective definition of blocking exists, complicating the comparison of different blocking studies. Furthermore, most common blocking indices involve thresholds tuned to observed variability: large apparent biases in GCM blocking climatologies can arise through small biases in the time-mean state (Doblas-Reyes et al., 2002). Pelly and Hoskins (2003a) emphasise the importance of longitude-dependent parameters in blocking indices for the accurate identification of blocking events.

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 coupled GCMs.

8.4.6 Atlantic Multidecadal Variability

The Atlantic Ocean exhibits considerable multidecadal variability with a period of about 50 to 100 years. This multidecadal variability appears to be a stable 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). Coupled models 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 coupled models originates from variations of the thermohaline circulation (THC). The mechanisms, however, that control the variations of the THC are quite different across the ensemble of coupled models. In most models, the variability can be understood as a damped oceanic eigenmode that is stochastically excited by the atmosphere. In a few other models, 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. The presence of the strong multidecadal variability in the Atlantic may mask any anthropogenic weakening of the THC for several decades (Latif et al., 2004; Knight et al., 2005).

8.4.7 El Niño-Southern Oscillation (ENSO)

The El Niño-Southern Oscillation (ENSO) phenomenon is the dominant mode of natural climate variability in the tropical Pacific on seasonal to interannual time scales. During the last decade there has been steady progress in simulating and predicting ENSO and the related global variability using coupled GCMs (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 amplitude of the SST anomalies in the eastern Pacific (AchutaRao and Sperber, 2005). 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 8.3.1) remains a major source of error in simulating the annual cycle in the tropics, 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 and typically have thermoclines that are far too diffuse (Davey et al., 2002). Most coupled GCMs fail to capture the meridional extent of the anomalies in the eastern Pacific and tend to produce anomalies that extent too far into the western tropical Pacific. Most, but not all, coupled GCMs 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, 2005) in that more models are consistent with the observed time scale for ENSO. Generally speaking, the models have too little low frequency variability (time scale longer then ENSO). Some of the weaknesses in the simulated amplitude and structure of the variability have been discussed in Davey (2002).

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 characteristic 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 breakthroughs are: (i) the recognition that forecasts must include quantitative information regarding uncertainty (i.e., probabilistic prediction) and that verification must include probabilistic measures of skill (Kirtman, 2003); and (ii) that a multi-model ensemble strategy may be the best current approach for adequately resolving forecast uncertainty (Palmer et al., 2004). Palmer et al. (2004, Figure 2), for example, 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, recent 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., 2004). 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 consists of large-scale regions of enhanced and suppressed convection (zonal wavenumbers 1–3) coupled to a deep-baroclinic, primarily zonal circulation anomaly. Together, they propagate slowly eastward (~5 ms^{-1}) along the equator from the western Indian Ocean to the central Pacific. The MJO is now appreciated to be an integral component of the tropical atmosphere-ocean climate system (e.g., Lau and Waliser, 2005; Zhang, 2005). It affects variability in both the Indian/Asian and Indonesian/Australian summer monsoons, impacting onset, break episodes, tropical cyclone development and mean monsoon strength. Interannual variation of MJO activity, while not necessarily predictable in southern summer (e.g., Hendon et al., 1999; Slingo et al., 1990) but possibly predictable in northern summer (Teng and Wang, 2003), constitutes a fundamental component of the interannual variation of these monsoons. The MJO, because of the slow eastward propagation of the associated surface heat flux and zonal stress anomalies across the western Pacific, interacts strongly with the evolution of El Niño/Southern Oscillation (ENSO; e.g., McPhaden, 1999).

Simulation of the MJO in both coupled and uncoupled climate models was (at the time of the TAR) and still remains unsatisfactory (e.g., Lin et al., 2005). In part, we are now demanding more of the climate 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 climate models at the time of the TAR were judged using gross metrics, e.g., evidence of a spectral peak at eastward zonal wavenumber one in the velocity potential (e.g., Slingo et al., 1996). The phase and spatial structure 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). Thus, while a model may simulate some large-scale characteristics of the MJO (e.g., a spectral peak at eastward wavenumbers 1–3 for periods 35–90 days in winds and precipitation), the simulation may be deemed unsuccessful when the detailed structure of the surface fluxes is examined (e.g., Hendon, 2000).

Contemporary coupled and uncoupled climate models are able to simulate a preponderance of eastward power compared to westward power at MJO time and space scales, especially in zonal wind but less so in convection. However, most models do not simulate a realistic spectral peak in the 40–50 day band that stands out above a red background spectrum (e.g., Lin et al., 2005).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 some models (e.g., Sperber et al., 2005), but this variability does not occur often enough or with sufficient strength so that the MJO stands out above the broad-band background variability. This under-simulation 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 the double ITCZs in the western Pacific and for erroneously strong convective signals to sometimes develop in the eastern Pacific ITCZ (e.g., Inness and Slingo, 2003).

Even though the MJO is probably not fundamentally a coupled mode (e.g., Waliser et al., 1999), 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 sea surface temperatures 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 coupled GCMs. 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 annual variation of MJO activity implies that the simulated MJO will improperly interact with climate phenomena that are tied to the annual cycle (e.g., the monsoons and ENSO).

Simulation of the MJO is also adversely affected by biases in the mean state. 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 eastward extent of convection associated with the MJO, thereby reducing the overall strength and coherence of the MJO (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 appears to be emerging that convective schemes based on local vertical stability that include some triggering threshold produce more realistic MJO variability than those based on moisture convergence and 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 Quasi-Biennial Oscillation (QBO) is a quasi-periodic wave-driven zonal-mean wind reversal that dominates the low-frequency variability of the lower equatorial stratosphere (10–100 hPa) and affects a variety of extratropical phenomena including the strength and stability of the wintertime polar vortex (e.g., Baldwin, et al., 2001). Recent efforts to model the QBO in GCMs that employ horizontal resolutions typical of climate-change studies have focused on wave driving by resolved waves (Takahashi 1996, 1999; Horinouchi and Yoden, 1998; Hamilton et al., 2001) and parameterized non-orographic gravity waves (Scaife et al., 2000; Giorgetta et al., 2002; McLandress, 2002).

The inability of resolved wave driving to induce a spontaneous QBO in climate models has been a notorious issue for some time (Boville and Randel, 1992; Hayashi and Golder, 1994; Hamilton et al., 1999). Only recently (Takahashi, 1996, 1996; 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 (e.g., moist-convective adjustment). 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 (GWD) to force a QBO has now been demonstrated by a number of studies (Scaife et al., 2000; Giorgetta et al., 2002; McLandress, 2002). A general requirement is the enhancement of input momentum flux in the tropics relative to that needed in the extratropics. The magnitude of this 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). At this time we require better observational estimates of deep convective variability to constrain parameterizations of deep convection, and in turn the amount of resolved tropical waves in climate models. This would allow a specification of input flux to non-orographic GWD schemes that is more realistic in terms of its magnitude and composition.

8.4.10 Monsoon Variability

The simulation of monsoon precipitation by GCMs has improved since the TAR but most models are still unable to simulate the inter-annual variation of rainfall accurately. In the second phase of the Atmospheric Model Intercomparison Project (AMIP II), simulations were performed with twenty different atmospheric GCMs using specified monthly mean SSTs for the period 1979–1995. Gadgil et al. (2005) examined the ability of these models to simulate the six extreme years (3 strong and 3 weak) in the Indian summer monsoon rainfall that occurred during the period 1979–1995. They showed that almost all the models were able to simulate the strong Indian monsoon of 1988 (associated with La Niña) but most models failed to simulate the strong Indian monsoon of 1994 (that was associated with large warming in the western equatorial Indian ocean). This indicates that most atmospheric GCMs capture the teleconnection between the equatorial Pacific and the Indian summer monsoon but not the linkage between the equatorial Indian Ocean and the Indian summer monsoon. Srinivasan (2003) has shown that errors in the simulation of tropical continental rainfall in the AMIP II GCMs is related to their inability to simulate correctly the relationship between rainfall and column water vapour. Liang et al. (2002) found no correlation between the ability of AMIP models to accurately simulate the annual cycle of rainfall in China and their ability to simulate monsoon interannual variability. Marengo et al. (2003) examined the tropical climate simulated by a version of the COLA model forced with globally observed SSTs for the period 1982–1991. They show that the inter-annual variability of rainfall is realistically simulated in Northeast Brazil, Amazonia, Central Chile, Southern Argentina–Uruguay, Eastern Africa, and the tropical Pacific regions. Held et al. (2005) show that two versions of the GFDL coupled GCM (CM2.0 and CM2.1) are able to simulate the decrease in rainfall in Sahel observed during the period 1950 to 1980. Cook and Vizy (2005) evaluated the simulation of 20th century climate in West Africa in the IPCC AR4 models. They found that the simulation of north Africa summer precipitation the IPCC AR4 climate models is not nearly as realistic as the simulation of summer precipitation over North America or Europe.

Ashrit et al. (2003) examined the simulation of the Indian monsoon in the CNRM coupled GCM and found that the model simulates the Indian summer monsoon well but overestimates winter precipitation. Semenov and Bengtsson (2002) evaluated the performance of the ECHAM4/OPYC3 coupled GCM. The model generally overestimates annual mean precipitation over the continents except for North Africa, India, the north- and south-eastern coasts of South America, and an area north of the Gulf of Mexico. Over the ocean the highest discrepancies were found in the tropical belt in those regions with the most intense precipitation. The model produces excessive precipitation in the tropical Indian Ocean and in the regions of the ITCZ and SPCZ: with less precipitation in the Indian and south-eastern Asia coasts and in the western equatorial Pacific. Lambert and Boer (2001) compared fifteen coupled GCMs that participated in the Coupled Model Inter-comparison Project (CMIP). They found large differences in the simulated and observed precipitation in the equatorial regions and in the Asian monsoon region.

8.4.11 Predictions Using "IPCC" Models

Here we focus on the few results of initial value climate 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 predictions

Climate model evaluation has traditionally been limited to monthly-mean output or monthly-mean statistics of higher frequency phenomena such as the diurnal cycle. However, since the TAR it has been shown that climate models can be integrated as weather prediction models if they are initialized with analyses from the latter (Phillips et al. 2004). This advance appears to be due to: (i) improvements in the weather prediction model analyses and (ii) increases in the climate model spatial resolutions. This has opened a fruitful new avenue to compare the output from climate models to observations from field experiments and to evaluate their prediction at the much shorter time scales that are characteristic of many physical processes such as cloud formation and cumulus convection. It is also beneficial in terms of tracing climate model biases to drifts in short-range forecasts from observed states (Pope and Stratton, 2002, Boyle et al., 2005), and in terms of the simulation of chemical or aerosols distributions which heretofore have primarily been studied in "offline" integrations driven by observed meteorology.

Seasonal predictions

A version of the HadCM3 (known as GloSea) coupled GCM has been comprehensively assessed for skill in predicting observed variations in seasonal climate out to a range of 6-months (see, e.g., Graham et al., 2005; Davey et al., 2002). 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. Graham et al. (2005) analysed 43 years of retrospective forecasts with the GloSea run from observed ocean-land-atmosphere initial conditions to a range of 6 months from four start dates each year. The integrations were performed in a 9-member ensemble. Key conclusions include: (i) six-month predicted and observed phases of ENSO, as represented by tropical Pacific SST, show good correlation; (ii) the model is able to reproduce the large-scale observed lagged responses to ENSO events in the tropical Atlantic and Indian Ocean SSTs; (iii) the model is capable of realistic prediction of anomaly patterns in North Atlantic SSTs, shown to have important links with the North Atlantic Oscillation (NAO) and seasonal temperature anomalies over Europe.

The Geophysical Fluid Dynamics Laboratory (GFDL) Seasonal-to-Interannual (SI) prediction model utilizes model version CM2.0 (Delworth et al., 2005) which is the same coupled GCM used in the IPCC assessments. Twelve month retrospective and contemporaneous forecasts were produced using an ensemble of six members. The forecasts were initialized starting from a global ocean data assimilation (Rosati, A. et al., 1997 and Derber and Rosati, 1989) using the ocean component of CM2.0 and observed atmosperic forcing combined with atmospheric initial conditions from the atmospheric component of the CM2.0 system forced with observed SSTs. The integrations were run from starting dates of January, April, July-December for 15 years starting in 1991. The 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 described in Anderson et al. (2004) were evident for the entire length of the forecast (12 months).

Decadal predictions
Smith et al. () report on a large set of 10-year ensemble hindcasts validated against observed climate variations since 1979. The hindcasts were carried out by initialising the HadCM3 coupled GCM with analyses of observed anomalies of the atmosphere and ocean state, and also included anthropogenic forcings (greenhouse gases and sulphate aerosol) based on the SRES B2 scenario. Natural forcings were specified by repeating the previous 11-year solar cycle and reducing volcanic aerosol exponentially with an e-folding time scale of one year. This approach captures, in principle, predictability arising from both internal climate variability (as in seasonal forecasting) and the response to external changes in radiative forcing (as in long-term coupled GCM projections initialised from pre-industrial conditions). Surface air temperature is predicted with significant skill throughout the 10-year period, both globally (Figure 8.4.1) and in many regions. Skill at lead times up to 2–3 years ahead arises mainly from the ability to predict internal variations (Figure 8.4.1a and 8.4.1b), while global skill beyond 7 years ahead arises exclusively from the warming trend in response to externally-forced climate change (Figure 8.4.1a and 8.4.1c). The ensemble spread (diagnosed from simulations distinguished by different starting conditions) increases with lead time (Figure 8.4.1c cf 8.4.1b), reflecting the loss of predictability associated with the chaotic growth of differences between ensemble members. Although events such as the 1997–1998 El Niño cannot be predicted a decade ahead, the ensemble spread is wide enough to indicate the possibility of such episodes (Figure 8.4.1c). However, the observed evolution lies outside the ensemble range more often than would be expected by chance, partly because major volcanic eruptions such as Mount Pinatubo are assumed not to be known in advance (note the warm bias of the hindcasts in Figure 8.4.1c from 1991–1995). Also, the ensembles do not yet sample modelling uncertainties (see Section 10.5). Development of this approach to include alternative model formulations could provide a basis for probabilistic climate projections on interannual to decadal (and possibly longer) time scales.

8.4.12 Summary

Since the TAR there has been progress in the representation of large-scale variability over a wide range of time-scales in coupled GCMs used for climate predictions of the future. Coupled GCMs capture the dominant extratropical patterns of variability known as the Northern and Southern Annular Modes (NAM and SAM), the Pacific Decadal Oscillation (PDO) and the Pacific-North American (PNA) and Cold Ocean-

Warm Land (COWL) Patterns. Coupled GCMs simulate Atlantic multidecadal variability although the relative roles of high and low latitude processes appear to differ from model to model. In the tropics, obtaining a completely accurate representation of the El Niño-Southern Oscillation (ENSO) and the Madden-Julian Oscillation (MJO) with coupled GCMs continues to present a challenge. Developments in model formulation since the TAR have generally led to improvements in the amplitude, structure and time-scale of these modes, yet systematic errors persist.

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. Moreover, 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.

Since the TAR, a large number of scientific papers have appeared which have analyzed the simulation of the extreme events in a variety of models. This is the first time that high-frequency (daily) data have been made available from a large number of models. For the economy of space, we will confine our report mainly to those papers which examine the simulation of extreme events in coupled models used for the IPCC assessment.

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. Based on this, a reasonable hypothesis might be that the extreme events are insensitive to global scale anthropogenic forcings. But that is not the case. Our assessment of the present scientific literature shows that the global statistics of the extreme events in the current climate, including the observed trends during the twentieth century, are well simulated by the current models. After making an extensive search of all the available scientific literature, we cannot find a single scientific report which shows that the observed trends in the extreme events during the twentieth century can be simulated without the anthropogenic forcing.

The successes of the AR4 models in simulating the extremes can be summarized by quoting directly from the scientific papers: "On the whole, the AGCMs appear to simulate temperature extremes reasonably well" (Kharin et al., 2004); "In agreement with observations, the models generally simulate modern cold air outbreaks most frequently over western North America and Europe, and least commonly over the Arctic" (Vavrus et al., 2005); "The model simulations agree with the observed pattern for late 20th century of a greater decrease of frost days in the west and southeast U.S. compared to the rest of the country, and almost no change in frost days in fall compared to relatively larger decreases in spring" (Meehl et al., 2005).

There has been little work to explore the impact of terrestrial processes in the capacity of climate models to simulate rainfall or temperature extremes. Pitman et al. (2004) and Bagnoud et al. (2005) used the AMIP-II methodology to explore whether the complexity of the land surface in climate models could affect the simulation of rainfall and temperature extremes. Bagnoud et al. (2005) showed that canopy interception was required in a land surface model to capture rainfall extremes, while surface tiling and a time-varying canopy conductance was needed to capture maximum temperature extremes. Most climate models used in the FAR include land surface models that explicitly model these processes. There is therefore no evidence that the capacity of climate models to simulate temperature and rainfall extremes is limited by uncertainty in how the terrestrial surface is modelled.

The assessment of extremes, especially for temperature, has been done in terms of the frequency, intensity or persistence of intense events. For precipitation, the assessment has been done either in terms of return values or extremely high rates of precipitation. In this section, we assess the extreme events 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.

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 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 explicit 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 National Center for Atmospheric Research/Department of Energy Parallel Climate Model (PCM). 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 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). Holt et al. () have compared extreme temperature indices in the NCEP reanalysis with that simulated by the HadCM3 GCM. They found that the HadCM3 GCM simulated the extreme temperatures well but was not as good as the regional HadRM3P model in the simulation of extremes.

Vavrus et al. (2005) 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 primary features of cold air outbreaks in the current climate with respect to location and magnitude.

Researchers have also established relationships between large scale circulation features and cold air outbreaks or heat waves. For example, Vavrus et al. (2005) 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 the 500hPa circulation pattern.

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 and Brown (2005) have shown that in the Hadley Center Global Model, although regional distributions of wet and dry area 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.

8.5.2 Extreme Precipitation

Since the TAR, many simulations with high resolution GCM have been made. Iorio et al. (2004) have 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. 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 mmday^{-1} was good but that exceeding 30 mmday^{-1} 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). Santos et al. () have shown that extreme winter precipitation events in Europe are related to the phase of the North Atlantic Oscillation. They have shown that the HadCM3 GCM is able to simulate the phase of the North Atlantic Oscillation, and hence will be useful to examine the changes in extreme winter precipitation in Europe in the future. Emori et al. (2005) have shown that a high-resolution AGCM 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%.

Using the Palmer Drought Severity Index (PDSI), Dai at al. (2004) concluded that very dry or wet areas (PDSI above +3 or below –3) have increased from 20% to 38% since 1972. Burke and Brown (2005) have shown that the Hadley Centre AGCM (HadAM3) is able to simulate this trend in PDSI only if the anthropogenic forcing is included in the 20th century simulation.

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 (typically T106 or higher) 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. 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 (typically ±2°C).

There is a substantial disagreement among the models about the effects of global warming on the intensity of tropical cyclones. Knutson and Tuleya (2004) find that “idealized hurricanes, simulated under warmer, high CO2 conditions, are more intense and have higher precipitation rates than under present-day conditions.” In contrast, Bengtsson et al. (2005) conclude that, “there are no indications in this study of more intense storms in the future climate, either in the tropics or extratropics…..” However, several idealized experiments with global warming SST show consistently that the frequency of tropical cyclones is reduced (Bengtsson et al., 2005;Yoshimura and Sugi, 2005).

There are but a few studies where a model’s ability to simulate tropical cyclones in the current climate has been analyzed in greater detail. Surprisingly, the current generation models have a remarkable ability to simulate the statistics and the geographical distributions of tropical cyclones. Bengtsson et al. (2005) have shown that the global metrics of tropical cyclones (tropical or hemispheric averages) are broadly reproduced by the ECHAM model, even as a function of intensity. Yoshimura and Sugi (2005) have also shown a realistic simulation of the geographical distribution of tropical cyclones.

Almost all the papers agree on one major result: the tracks of tropical cyclones are affected by the structure of the tropical SST in any given year (viz. El Niño vs. La Niña), and models are able to simulate those differences. This is especially relevant to the impact on society, because changes in the tracks of destructive cyclones can be as important (or even more important if hurricanes pass over highly developed population centers) as the changes in the intensity. Observational studies have shown systematic shifts in the tracks of western North Pacific typhoons during the past 50 years. However, there are no comparable modeling studies to assess the causes of changes in the tracks in the twentieth century.

8.5.4 Summary

Because coupled models have coarse resolution and large 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. This is especially true for the temperature and wind-related extremes. There is no agreement among the models whether global warming will make tropical cyclones more or less intense. There seems to be some agreement among the models that the frequency of tropical cyclones will be reduced. Models continue to show serious deficiencies in the simulation of precipitation, both in the intensity and the distribution of precipitation. As long as climate models do not have sufficient resolution to explicitly resolve at least the large convective systems and must use parameterizations for deep convection, it is unlikely that simulation of precipitation will be satisfactory.

8.6 Climate Sensitivity and Feedbacks

8.6.1 Introduction

The concept of climate sensitivity, which is broadly defined as the equilibrium global mean surface temperature change following a doubling of atmospheric CO_2 concentration, is being used to characterize the response of the global climate system to a given forcing. Throughout the four IPCC assessments, the range of climate sensitivity estimates from climate models plays a central role in the discussions of uncertainty associated with projections of future climate change (Chapter 10). This range results mostly from differences among models in the way internal feedback processes amplify or damp 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 includes the Last Glacial Maximum (Chapter 6), and the evolution of climate over the last millenium and the 20th century (Chapter 9). The compilation and the comparison of climate sensitivity estimates derived from models and from observations are presented in Chapter 10 (Box10.2). 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.

We assess below the reasons why the estimates of climate sensitivity and of climate feedbacks differ among current models (8.6.2), 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 the reliability of these processes in the global climate models used to make projections of future climate change (8.6.3). Finally we discuss how we may assess our relative confidence in the different climate sensitivity estimates derived from climate models (8.6.4).

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 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 (these ocean models are usually refered 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) (Cubash et al., 2001) is defined as the globally averaged surface air temperature difference for the 20-year period around the time of CO_2 doubling minus the control run. That response depends both on the sensitivity and on the ocean heat uptake. To link the equilibrium climate sensitivity and the transient climate response, an *effective climate sensitivity* has been defined (Murphy 1995) in transient climate change integrations. It is computed from the oceanic heat storage, the radiative forcing and the surface temperature change (Cubash et al., 2001; Gregory et al., 2002).

The climate sensitivity estimate 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, 2003c). The global annual mean surface temperature change 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 gives no indication of the response of any climate variable other than surface temperature, nor of the occurrence of abrupt changes or extreme events. Despite its limitations, the climate sensitivity constitutes however 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 to 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 variability.

8.6.2.2 Why have the model estimates changed since the TAR ?

Most climate models have undergone substantial developments since the TAR (and probably more than between the SAR and the TAR), that generally consist in improved parameterization of specific processes such as clouds, boundary layer or convection (Section 8.2). In some cases, developments concerned also numerics, dynamical core or the coupling to a new component (ocean, carbon cycle, etc.). Developing new versions of a model so as to improve the simulation of the current climate is at the heart of modelling group activities. The rationales for these changes are generally based on a combination of process-level tests against observations or against cloud-resolving models or large-eddy-simulation models (Section 8.2), and of the quality of the overall simulation of the model (Sections 8.3 and 8.4). Climate sensitivity estimates are generally not part of the decision process of making such or such change in the model. However, developments can, and do, affect the climate sensitivity estimate of models.

The climate sensitivity estimate from the latest model version used by modelling groups has increased (e.g., CCCma/CGCM, NCAR/CCSM, MPI/ECHAM, MRI), decreased (e.g., CCSR/NIES, GFDL) or remained unchanged (e.g., IPSL, Hadley Centre) compared to the TAR. In some models, changes in climate sensitivity are ascribed primarily to changes in the cloud parameterization or in the representation of cloud-radiative properties (e.g., CGCM, CCSM, MRI, CCSR), or to changes in the planetary boundary-layer and sea-ice (e.g., GFDL). However, in most models the change in climate sensitivity cannot be attributed to a specific change in the model. For instance, Williams et al. (2005b) 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 of change B does not produce the same as the change A+B. 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. For instance, the introduction of the Lock boundary layer scheme (Lock et al., 2000) to HadCM3 has a minimal impact on the climate sensitivity, in contrast to the introduction of the scheme to the GFDL model (Soden et al., 2004; Williams et al., 2005b).

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. Webb et al. (2005) show that differences in the models' feedbacks contribute approximately three times more to the range of equilibrium climate sensitivity estimates than differences in the models' radiative forcing (the spread of models' forcing is discussed in 10.2). 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., 2005) and in transient climate change integrations (Soden and Held, 2005).

Several methods have been used to diagnose climate feedbacks in GCMs, whose strengths and weaknesses are reviewed in Bony et al. (2005). Whatever the approach being used, the partial radiative perturbation (PRP) or radiative-convective method (RCM) analysis (Colman, 2003a), a variant of the PRP analysis (Soden and Held, 2005), or the CRF approach (Webb et al., 2005), all studies suggest that the spread of models' feedbacks primarily stems from the large range of *cloud* radiative feedbacks (Figure 8.6.1). Cloud feedbacks, whose sign and range are discussed in 8.6.3.2.2, therefore constitute the largest source of uncertainty in current model estimates of climate sensitivity.

[INSERT FIGURE 8.6.1 HERE]

The water vapour feedback (discussed in 8.6.3.1) constitutes a strong positive feedback in climate models. A substantial spread is noticed in the strength of this feedback, larger in Colman (2003a) than in Soden and Held (2005). It is not known whether this indicates a closer consensus among current OAGCMs than among older models, differences in the PRP (or RCM) methodology, or differences in the nature of climate change integrations among 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, 2005). Since relative humidity (RH) is nearly unchanged (Section 8.6.3.1), temperature and specific humidity changes are highly correlated in climate change. As a result, the water vapor and lapse rate feedbacks have a degree of anti-correlation, and this makes intermodel differences in the combination of water vapor and lapse rate feedbacks a substantially smaller contributor to the spread in climate sensitivity estimates than differences in cloud feedback (Figure 8.6.1). 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, 2005).

The global surface albedo feedback associated with snow and sea-ice changes has been estimated using different methodologies (Colman, 2003a; Soden and Held, 2005; Winton, 2005). All three studies suggest that it is positive in all the models, substantially weaker than the water vapour feedback, and that its range among models is much smaller than that of cloud feedbacks. Winton (2005) suggests that about three-quarters of the global feedback arises from the Northern Hemisphere (8.6.3.3).

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 and cloud feedbacks separately. Although this division can be regarded as somewhat artificial because 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 address separately the water vapour/lapse rate feedbacks and then the cloud and surface albedo feedbacks.

Note that feedbacks associated with the carbon cycle are not discussed in this section; carbon feedbacks affect the rate of CO_2 increase in the atmosphere but do not affect climate sensitivity, which is defined with respect to a specified CO_2 forcing (e.g., a CO_2 doubling).

8.6.3.1 Water vapour and lapse rate

Absorption of LW radiation increases approximately with the logarithm of water-vapour concentration, and the Clausius-Clapeyron equation dictates a near-exponential increase in moisture holding capacity with temperature. Combined, these constraints predict a strongly positive water vapour feedback if RH is close to unchanged. To a first approximation, GCMs do 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, 2005; Figure 8.6.1).

In the PBL, humidity is controlled by strong coupling with the surface, and a broad-scale quasi-unchanged RH response is uncontroversial (Wentz and Schabel, 2000; Trenberth et al., 2005). For the extratropics, confidence in GCMs' water vapour feedback is also relatively high 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., 2005 for a review).

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 LW cooling, and changes in their area or humidity could potentially have a significant impact on 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 confidence in water vapour feedback.

In contrast to cloud feedback, a strong positive water vapour feedback is a robust feature of GCMs (Stocker et al., 2001; Section 8.6.3), being found across models with many different schemes for advection, convection and condensation of water vapour. High resolution mesoscale (Larson and Hartmann, 2003a,b) 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, also under idealized warming, 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 some 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. The TAR noted the 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 rather 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 to 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).

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, unless considered over entire circulation systems (e.g., Lau et al., 1996). 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. 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). 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 feedbacks and 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. 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).

Major features of the mean humidity distribution are reasonably simulated in GCMs, along with the consequent distribution of OLR (Section 8.3.1). In the important subtropical subsidence regions, models show a range of skill in representing the mean UTH (Allan et al., 2003; Chung et al,. 2004; Brogniez et al., 2005). 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 (Brogniez et al., 2005). 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 that given the near-logarithmic dependence of LW radiation on humidity, errors in the control climate humidity have little *direct* effect on climate sensitivity: it is the fractional change of RH 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. Allan et al. (2003) found an AGCM forced by observed SSTs simulated interannual changes in tropical mean simulated 6.7μm radiance (sensitive to UTRH and temperature) in broad agreement with HIRS observations over the last two decades. Minschwaner et al. (2005) analysed the interannual response of tropical mean 250 hPa RH to the mean SST of the most convectively active region in 16 AR4 CGCMs. 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).

One test of the response of free tropospheric 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 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. Caution is required, however, 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 CGCM study has found similar global LW 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. 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 (Figure 8.6.2), 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 W $m^{-2}K^{-1}$, a range which covers that of models under GHG forcing. Using estimated aerosol forcing, Soden et al. (2002) found a model simulated response of HIRS 6.7μm 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). An important caveat on these studies is that climate perturbation from Pinatubo is small, not sitting clearly above natural variability (Forster and Collins, 2004).

[INSERT FIGURE 8.6.2 HERE]

Indirect supporting evidence for model water vapour feedback strength also comes from paleo modeling, where models have had some success in simulating paleo climates under CO_2 and other forcing (Section 6.4.2.1). Without strong positive water vapour feedback, models would have difficulty in reproducing the size of changes such as the tropical cooling in the last glacial maximum (Pierrehumbert, 1999). Other indirect supporting evidence for feedback strength is that suppressing humidity variation from the radiation code in a CGCM produces unrealistically low interannual variability (Hall and Manabe, 1999).

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, 2005). There has been extensive testing of GCM tropospheric temperature response against observational trends for climate change detection purposes (section 9.4.4). Although some recent studies have suggested consistency between modelled and observed changes (e.g., Fu et al., 2004), 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 (Allan et al., 2002a; Colman, 2003a).

In the stratosphere, GCMs' 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 strong observational evidence that stratospheric water vapour increases with temperatures near the tropopause on seasonal (Mote et al., 1996) and interannual (Randel et al., 2004) timescales. On longer timescales, however, the link is less clear, due to uncertainties in the magnitude and source of humidity trends, and their association with observed temperature changes (Section 3.4.2.4). Transport processes between the troposphere and stratosphere remain the subject of debate (e.g. Sherwood and Dessler, 2001; Rosenlof, 2003), but if the long-term lower stratospheric trend measured at Boulder is representative of global trends (Section 3.4.2.4) and is predominantly a feedback response, this would imply a strong 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. Although a degree of spread in lapse rate-water vapour feedback is apparant between GCMs, no substantial evidence suggests that the broadscale RH response of models to climate change constitutes an artefact of GCMs. Indeed, new evidence from both observations and models has reinforced the traditional 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.

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 SW and 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 net radiative effect of clouds on climate may change and thereby produce a radiative feedback on climate warming (Randall, 2000; NRC, 2003; Zhang, 2004; Stephens, 2005; Bony et al., 2005).

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; Ogura et al., 2005; Yokohata et al., 2005; Zhang, 2004). Moreover, the spread of climate sensitivity estimates among current models primarily arises from inter-model differences in cloud feedbacks (Colman, 2003; Soden and Held, 2005; Webb et al., 2005; Section 8.6.2, Figure 8.6.1). Therefore, cloud feedbacks still constitute 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 (8.6.3.2.1), in the interpretation of the range of cloud feedback estimates among current climate models (8.6.3.2.2), and in evaluation of the model cloud feedbacks using observations (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 to understand how a change in climate may affect the spectrum and the radiative properties of these different clouds, and to estimate 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. (2005). 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 might be independent on surface temperature, which would represent an important constraint on tropical cloud-climate feedback. 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 in lockstep with the temperature (Tompkins and Craig, 1998). On the other hand, the response of the anvil cloud fraction to a change in temperature remains an object of debate. Assuming that the 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; Dessler and Minshwaner, 2005), leading to a vigorous debate with the authors of the hypothesis (Bell et al., 2002; Chou et al., 2002; Lindzen et al., 2002). Another observational study (Del Genio and Kovari, 2002) suggests 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, 1998). 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 into 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's measure, that can predict boundary-layer cloud amount as well as the Klein and Hartmann (1993)'s measure, would not necessarily predict an increase in low-level clouds in a warmer climate (Williams et al., 2005; Wood, 2005). 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 et al., 1992; 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 a prevailing thick, high-top frontal clouds in regions of synoptic ascent and low-level 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 (2005) show that the decrease in storm frequency has a larger radiative impact than the increase in storm intensity, and suggest that this would produce decreased reflection of SW radiation and a positive anomaly of a few W/m^2 on the net radiation budget. The poleward shift of the storm tracks may further decrease the amount of SW radiation reflected (Tsushima et al., 2005a). 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, 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. Although these different studies suggest the potential for a decreased cooling effect of extratropical clouds in a warmer climate, further studies are required to confirm the sign and calculate the magnitude of extratropical cloud feedbacks in climate change.

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 (Webb et al., 2005; Soden and Held, 2005). Several studies suggest that the sign of cloud feedbacks may not be necessarily that of CRF changes (Zhang et al., 1994; Colman, 2003; 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 PRP method, that excludes clear-sky changes from the definition of cloud feedbacks, diagnoses a positive cloud feedback in virtually all the models (Colman, 2003; Soden and Held, 2005). However, the cloud feedback estimates diagnosed from either the change in CRF or the PRP method are well correlated, and they exhibit a similar range among models.

By decomposing the model feedbacks into regional components (Boer and Yu, 2003; Volodin, 2004; Stowasser et al., 2005; Webb et al., 2005; Williams et al., 2005) or dynamical regimes (Bony et al., 2004; Bony and Dufresne, 2005; Wyant et al., 2005), substantial progress has been made in the interpretation of this spread. The current intermodel difference in global cloud feedbacks arises mostly from the shortwave response of clouds in the tropics (Figure 8.6.3; Webb et al., 2005). Several studies have analyzed the diversity of tropical cloud responses to climate warming simulated by GCMs. The comparison of coupled ocean-atmosphere models used for the climate projections of chapter 10 (Bony and Dufresne, 2005), of atmospheric or slab ocean versions of current models (Webb et al., 2005; Wyant et al., 2005), or of slightly older models (Bony et al., 2004; Volodin, 2004; Stowasser et al.; 2005) all suggest a dominant role for boundary-layer clouds in the diversity of tropical cloud feedbacks, and highlight the importance, for the overall CRF response, of the clouds response in subsidence regimes (Figure 8.6.4). Models also predict different responses of deep convective clouds, but these differences contribute comparatively less to the diversity of model cloud feedbacks. In middle latitudes, differences in the representation of mixed-phase clouds and in the degree of latitudinal shift of the storm tracks predicted by the models contribute to intermodel differences in the extratropical shortwave CRF response to climate change (Tsushima et al., 2005a; Ogura et al., 2005). The contribution of polar cloud feedbacks to the range of global cloud feedbacks is currently unknown.

[INSERT FIGURE 8.6.3 HERE]

[INSERT FIGURE 8.6.4 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. 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; Lin and Zhang, 2004; Weare, 2004; Williams et al., 2003; Wyant et al., 2005), and has thus become more constraining. 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; Gordon et al., 2005; Williams et al., 2005; Wyant et al., 2005). An observational test focused on the global response of clouds to seasonal variations has been proposed to evaluate model cloud feedbacks (Tsushima et al., 2005b), but it has not been applied to current models yet.

These studies highlight some common biases in the simulation of clouds by current models. This includes the overprediction of optically thick clouds and the underprediction of optically thin low and middle-top clouds. Although these errors may eventually compensate and lead to a prediction of the mean CRF in agreement with observations (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.

As an attempt to assess some components of the clouds 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 condition. When examining atmosphere-mixed-layer ocean models, Williams et al. (2005) 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. Stowasser and Hamilton (2005) and Bony and Dufresne (2005) have examined the ability of the OAGCMs of Chapter 10 to simulate the change in tropical CRF to a change in large-scale vertical velocity, in lower tropospheric relative humidity, and in sea surface temperature. They show that the models exhibit the largest diversity and the largest errors vis-a-vis observations 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.

In the middle latitudes, several processes would need to be evaluated in climate models to assess the reliability of the models cloud feedbacks (Section 8.6.3.2.1). Such evaluations have been carried out for only a few models. For instance, Tselioudis and Rossow (2005) show that the GISS model captures but overestimates the radiation anomalies associated with a change in the frequency or the intensity of extratropical synoptic systems. Also, modelling assumptions controlling the cloud water phase (liquid, ice or mixed) are known to have a substantial impact on the prediction of extratropical cloud feedbacks (e.g.. Ogura et al., 2005; Tsushima et al., 2005). But few evaluations of the assumptions used in current models are available. Therefore, it is too early to assess the reliability of the enhanced cooling effect of extratropical clouds predicted by models in response to climate warming (Figure 8.6.3).

8.6.3.2.4 Conclusion on cloud feedbacks

Despite some advances in our understanding of the physical processes that control the clouds' 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, much 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 models. This is likely to foster more specific observational analyses and model evaluations, that will improve future assessments of cloud feedbacks.

8.6.3.4 Cryosphere feedbacks

A number of feedbacks that significantly contribute to the global climate sensitivity are introduced by the cryosphere. A robust feature of the response of climate models to increases in atmospheric concentrations of greenhouse gases 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 any considerable convergence in the latest generation of AOGCMs (Chapman and Walsh, 2005; see also Section 11.3.8). 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

TAR, some progress has been demonstrated 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 (2005) suggest that the performance of the AR4 models in reproducing the observed seasonal cycle of the land snow cover (especially the springtime melt) constitutes an indirect evaluation of the snow-albedo feedback simulated by the models in climate change scenarios, and so may provide a constraint that would reduce the divergence in simulations of snow albedo feedback (Figure 8.6.5). They found that the feedback has a pronounced interhemispheric asymmetry and the relative contributions of snow and sea ice to the enhanced simulated warming differ dramatically between the southern and northern hemispheres. In the northern hemisphere, the simulated annual mean increase in solar radiation resulting from the shrunken cryosphere has almost equal contributions from snow and sea-ice retreat, while in the southern hemisphere the relative contribution of terrestrial snow to the polar amplification is nearly negligible. A new result found independently by Winton (2005) 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 8.6.5 HERE]

Our understanding of numerous 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., 2005) . 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 TAR, particularly by the inclusion of more sophisticated dynamics by most of them (Section 8.3.3), 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 AGCM/UML (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) showed 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.

While playing the central role in polar amplification, the cryosphere feedbacks are likely to be not the only ones. Recent studies (Alexeev et al., 2003; Alexeev et al., 2005; Cai, 2005) suggest that feedbacks associated with atmospheric dynamics which are not directly dependent on the cryosphere can contribute to polar amplification. However, the present day understanding of processes controlling polar amplification and the interdependence of those processes is insufficient to allow their contributions to be quantified.

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

To better assess our relative confidence in climate projections from the different models, and at least constrain their range among climate models, one would need to apply to all the models a similar and large set of observational tests (i) allowing us to *measure* the deviation between simulations and observations, and (ii) *discrimitating* for model estimates of different characteristics of climate change: global climate warming, large-scale patterns of climate change (interhemispheric symetry, polar amplification, vertical patterns of temperature change, land-sea contrasts), regional patterns, transient aspects of climate change, etc. Such an ensemble of tests is referred to as “*climate metrics*”. To guarantee the robustness of the metrics, it would be necessary to use, as much as possible, robust and independent sets of observations, and different methodologies for model-data comparisons. Specific climate metrics could be developed to assess these different characteristics. For example, assessing our confidence in model projections of the Australian climate would require a set of observational tests including at least some diagnostics related to the simulation of ENSO because the Australian climate depends much on it (11.3.7.1).

To better weight our confidence in the different model estimates of climate sensitivity, one may apply two kinds of observational tests to climate models: tests related to the global climate response associated with specified external forcings (this is discussed in Chapters 6 and 9), and tests focused on the simulation of key feedback processes.

Based on our understanding of the physical processes that control key climate feedbacks (8.6.3), and of the origin of intermodel differences in the simulation of feedbacks (8.6.2), some necessary (although probably not sufficient) processes should be considered as part of a metrics focused on climate feedback processes. These processes include, for the water vapor and lapse rate feedbacks: the response of upper relative humidity and lapse rate to interannual or decadal changes in climate; 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; for snow-albedo feedbacks: the relationship between surface air temperature and snow melt over northern land areas during springtime; for sea-ice feedbacks, the simulation of sea-ice thickness.

A number of such diagnostic tests have been proposed since the TAR (8.6.3). However, very few of them have been applied to a large fraction of the models currently in use. Moreover, for specific and critical aspects of climate change feedbacks (e.g. the clouds response to a change in environmental conditions), one would need diagnostic tests using different observational datasets, different methodologies, and different timescales to reach robust conclusions. Therefore, it is too early to use any particular metrics to weight our relative confidence in different climate feedback estimates from current models.

8.7 Mechanisms Producing Thresholds and Abrupt Climate Change

8.7.1 Introduction

Before beginning the discussion of thresholds and abrupt climate change, one must define what is meant by “threshold” and “abrupt”. Here we use the definitions put forth in “Abrupt Climate Change: Inevitable Surprises” (reference?). The climate system tends to respond to changes in a gradual way until it crosses some threshold. At this threshold, the climate system responds to forcing changes in a nonlinear way. That is, over some time period 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 definition 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 find 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 seem worthy of some discussion here.

8.7.2 Forced Response

8.7.2.1 Thermohaline circulation changes

As the radiative forcing of the planet changes, the climate system responds on many different time scales. For the physical climate system (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 thermohaline circulation (see Box 5.1 in Chapter 5 for definition and description) is a main area of study.

The meridional overturning circulation (MOC) transports large amounts of heat (order of 10^{15} watts) and salt into high latitudes of the N 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. Both paleo-studies (e.g., Broecker 1997, 2000) 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. Some modeling studies (Rahmstorf, 1995; Tziperman, 1997; Rind et al., 2001) suggest that there are thresholds where the THC may suddenly weaken or even shut down causing abrupt climate changes.

It is important to note in this discussion 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 than the equilibrium response. This 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). 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 recovers to its unperturbed value.

In most (but not all) AOGCMs, the MOC weakens as the climate warms (see Chapter 10 discussion). The amount of the weakening varies from model to model. As the MOC weakens, 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 N 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.

Some researchers have speculated that the change of state of the MOC (on vs. off) could produce changes large enough to cool to the Northern Hemisphere as GHG increase and potentially cause a future ice age (e.g.. Joyce and Kegwin. 2004). However, no AOGCM has supported this speculation when forced with realistic estimates of future climate forcings (see more discussion on this topic in Chapter 10). In addition, 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., 2005).

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 in the climate system on decadal to centennial time scales. Idealized studies 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 vary widely among models (Stouffer et al., 2005) which lead 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).

In addition to the magnitude of the freshwater input, the exact location may also be important (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. It seems likely that models can produce reliable forecasts of THC behavior over the next century or so in response to changes in the GHG forcing, however the reliability of longer term forecasts is unknown.

The processes determining MOC response have been studied in a number of models. In many models, initial MOC response to increasing greenhouse gases 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; see Chapter 10). More complex feedbacks, associated with wind and hydrological changes, are 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; Gregory et al., 2003; Hu et al., 2004). The magnitudes of the climate factors causing the THC to weaken, the feedbacks and the associated restoring factors are 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; see also Chapter 9). Model intercomparison studies (Gregory et al., 2005; Stouffer et al., 2005) were developed to identify and understand the causes for the wide range of THC responses in the AR4 models (Chapter 10).

8.7.2.2 Rapid West Antarctic and/or Greenland ice sheet collapse

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). 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, the reader is encouraged to 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 THC (8.7.2.1).

First experiments with three-dimensional ice sheet models coupled to AOGCMs indeed show the possibility of such behavior. In the study by Fichefet et al. (2003), enhanced freshwater input from increased melting of the Greenland ice sheet causes an abrupt weakening of the THC of about 4 Sv by the end of the 21st century under an average climatic warming scenario. In this experiment, the additional freshwater input from the Greenland ice sheet peaked at about 0.03 Sv, enough to induce significant a cooling over eastern Greenland and the northern North Atlantic. This cooling tends to stabilise the Greenland ice sheet by reducing melting rates to present day values, and weakened the initial warming over northwestern Europe and most of Canada by 1 to 3°C for at least a decade. Another experiment with the same ice-sheet model fully two-way coupled with HadCM3 under constant $4 \times CO_2$ idealized forcing shows an initial peak freshwater flux in the Atlantic of 0.06 Sv as compared to a non-coupled run, which causes a temporary 1–2 Sv decline in the THC. However, the circulation fully recovers after 300 years. The weaker coupling between ice-sheet and climate reflects the sensitivity range of the oceanic component of AOGCMs as well as differences in locations of main sites of deep convection and meltwater input.

On longer time scales, for a sustained summer warming in excess of 10°C, complete melting of the Greenland ice sheet could take as little as 1000 years (Gregory et al., 2004, and references therein). In that case, meltwater discharge would peak at between 0.2 and 0.3 Sv, enough to pass the threshold for major weakening of the THC in most ocean models or even halt the THC in others (Rahmstorf, 1995; Stouffer et al., 2005).

The potential disintegration of the Greenland ice sheet over the third millennium could affect the atmospheric circulation because of reduced atmospheric blocking. Model studies with HadCM3 find a winter cooling over Scandinavia and the western Arctic together with a northward shift of precipitation patterns over Greenland due to changes in storm tracks and interaction with sea ice (Toniazzo et al., 2004). Similar findings are reported by Lunt et al. () with the French IPSLCM4 AOGCM.

The response of the Atlantic THC to changes in the Antarctic ice sheet is less understood. Experiments with ocean-only models where the meltwater changes are imposed as surface salinity changes, indicate that the Atlantic THC will intensify as the waters around Antarctica become lighter (Seidov et al., 2001). 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, leading to a weakening of the Atlantic THC. In both model results, the Southern Hemisphere THC associated with Antarctic bottom water formation weakened, causing a cooling around Antarctica.

Although there is a clear potential for increased Antarctic fresh water input from increased melting of ice shelves and icebergs (Marsland and Wolff, 2001; Williams et al., 2001; Beckmann and Goosse, 2003; Shepherd et al., 2003), and an increased flux of ice across grounding lines (Thomas et al., 2004), total fresh water volumes are likely to be significantly lower than for Greenland. In addition, the freshwater would be spread out over a much larger area, leading to a lower local rate of freshening of surface waters (Stouffer et al., 2005).

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 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 (less than 3 years or so). 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 (Glecker et al., 2005; Delworth and Stenchikov, 2005).

Modeling studies of large volcanoes suggest that it is extremely unlikely that volcanic eruptions could produce enough cooling to overcome the projected warming over the next century (Bertrand et al., 2002). A modeling study of a super volcano suggests that the cooling which lasts a few decades is not large enough to trigger a stadial period.

The models' ability to simulate the response of the climate system to volcanic eruptions is similar to their ability to simulate the climate response to future changes in GHG. Both produce changes in the radiative forcing of the planet.

8.7.2.3 Methane hydrate instability/permafrost methane

Methane hydrates are stored in the oceans along continental margins where they are stabilized by in situ water pressure and temperature fields, implying that ocean warming may cause hydrate instability and release of methane into the atmosphere, leading to further warming. Methane is also stored in the soils in areas of permafrost. Again, warming increases the likelihood of a positive feedback in the climate system. The warming would lead to increased permafrost melting, which releases the trapped methane into the atmosphere. The likelihood of potential future releases of methane from either methane hydrates found in the oceans or methane trapped in permafrost layers is assessed in Chapter 7.

Here we consider the potential for those releases to trigger 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. Since these changes produce changes in the radiative forcing through changes in the GHG concentrations, the climatic pacts 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 the changes should be similar to their ability to simulate future climate changes due to changes in the GHG forcing and any associated abrupt climate changes.

8.7.2.4 Biogeochemical

There are two aspects to this question. One is can biogeochemical changes lead to abrupt climate change? The second aspect is if abrupt changes in the THC can further impact the radiative forcing through biogeochemical feedbacks?

Abrupt changes in biogeochemical systems, of relevance to our capacity to simulate the climate of the 21^{st} 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) remain significant uncertainties. In part this is due to lack of understanding of processes (see Friedlingstein et al., 2003) 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., submitted).

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

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

The models' ability to simulate the response of the climate system to changes in the biogeochemical system is similar to their ability to simulate the climate response to future changes in GHG. Both produce changes in the radiative forcing of the planet. The ability of the models to simulate abrupt changes in the THC is discussed in Section 8.7.2.1.

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 as outlined above. 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 bi (or more) 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.

One interesting example of this case of abrupt climate change is found in Hall and Stouffer (2001). In an ultra-long AOGCM control integration (15,000 model years), they found 2 cases of large, abrupt climate events. In the North Atlantic event which they describe, the surface air temperature falls more than 10 standard deviations from the mean for a period of 15 to 20 years in response to a 4 standard deviation wind anomaly – a very non-linear response. The anomalous cold surface temperatures extend across the whole of the N Atlantic and into West Europe. However, due to a dynamical atmospheric response to the N Atlantic SST anomaly, the middle latitude continents and the hemispheric mean temperature are slightly warmer than normal during the event.

Similar events caused by a spontaneous transition to an infrequently visited state have been found in other climate models (e.g., Goosse et al., 2002). Again, these events are associated with changes in the ocean circulation, mainly in the N Atlantic. The event can last for several years to a few centuries. They bear some similarities with the conditions observed during relatively cold period in the recent past in the Arctic (Goosse et al., 2003)

Unfortunately, the probability to have such an event is difficult to estimate as it is requires a very long experiment and is certainly dependant on the mean state simulated by the model. Furthermore, comparison with observations is nearly impossible since it would require a very long periods with constant forcing which do 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 the 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 hierarchy of models of differing levels of complexity, each of which 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 coupled GCMs. These models, which include more and more components of the climate system (e.g., Fichefet et al., 2003; Friedlingstein et al., 2003), are designed to provide the best representation of the system and its dynamics, thereby serving as the most realistic laboratory of nature. In particular, they describe many details of the atmospheric and oceanic flow patterns, such as individual weather systems and regional oceanic currents. They are therefore the only modelling tools capable of simulating realistically the natural climate variability, extreme events and climate change feedbacks at both global and regional scales. Their major limitation is their high computational cost. Even using the most powerful computers, only a limited number of multi-decadal experiments can be performed with such models, 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 (e.g., Harvey et al., 1997). 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 coupled GCMs 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 coupled GCMs because of their computational expense. A particularity of simple climate models is that climate sensitivity and other subsystem properties must be specified based on the results of coupled GCMs or observations. Therefore, simple climate models can be tuned to individual coupled GCMs 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 coupled GCMs and simple climate models, Earth system models of intermediate complexity (EMICs) have been proposed (Claussen, 2000; Claussen et al., 2002; McGuffie and Henderson-Sellers, 2005). Given that this gap is quite large, there is a wide range of EMICs. 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 coupled GCMs, albeit in a more parameterised form. They explicitly simulate the interactions between various components of the climate system. Similarly to coupled GCMs, 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 designed for this purpose. Moreover, like simple climate models, EMICs can explore the parameter space with some completeness and are thus suitable for assessing uncertainty. EMICs can also be used 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 coupled GCMs. 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 utilising a statistical-dynamical approach. A priori, it is not obvious how the reduction in resolution or dynamics/physics affects the simulated climate. As shown below in Section 8.8.3, at large scale, most EMIC results compare favourably against observational data and coupled GCM 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 the AR4 to emulate the projections of future climate change conducted with state-of-the-art coupled GCMs, 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. Where possible, uncertainties in the forcing and in the feedbacks related to the carbon cycle are carried forward into the atmosphere-ocean module.

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 in the vertical direction with inter-hemispheric exchange in the mixed layer. In addition to the seven tuned model versions used in the TAR, the model is being tuned to outputs from the IPCC AR4 coupled GCMs driven by a 1% compound increase in CO_2 concentration (see www.pcmdi.llnl.gov/ipcc_for_analysts.php for information on IPCC AR4 model ouputs).

Data availablility has allowed four models from the IPCC AR4 coupled GCM dataset to be tuned to date. The procedure followed is similar to that described in the TAR (see Appendix 9.A), and the parameter values are given in Table 8.8.1. The first step is to select appropriate values for the radiative forcing for a CO_2 doubling, F_{2x} (W m^{-2}), and the climate sensitivity, T_{2x} (°C), which are two key parameters of the simple climate model. F_{2x} is fixed to the value supplied by the coupled GCM modelling group, and $T_{2x,}$ is set equal to the effective climate sensitivity of the coupled GCM (Raper et al., 2002). As well as the global mean surface temperature change, attempts are made to match both the land and ocean surface temperature changes with coupled GCM results by adjusting the equilibrium land-ocean sensitivity ratio and the land-ocean and inter-hemispheric heat exchange rates. The rate of change of the upwelling velocity is parameterised as a function of temperature change. The tuning process then consists of matching the coupled GCM net heat flux across the ocean surface by adjusting the ocean effective vertical diffusivity.

The linked modules of the simple climate model enable the mapping out and concatenation of uncertainties in emission scenarios, carbon cycle-related feedbacks, radiative forcing and climate models.

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, 2000; Claussen et al., 2002): integration, i.e., the number of interacting components of the Earth's climate system being explicitly represented in the model (hence the term integration is employed here in the sense of integrated modelling rather than in its original mathematical meaning), 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 8.8.2. A comprehensive description of all EMICs in operation can be found in Claussen (2005) and is available on the web via www.pik-potsdam.de/emics. 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 is reduced for the sake of enhancing integration, i.e., the simulation of feedbacks between as many components of the climate system as feasible. Others, with a lesser degree of integration, are utilised for long-term ensemble simulations to study specific aspects of climate variability. The gap between some of the most complicated EMICs and coupled GCMs is not large. Actually, this particular class of EMICs is derived from coupled GCMs. On the other hand, EMICs and simple climate models differ much more. This reflects the notion that EMICs as well as coupled GCMs tend to preserve the geographical integrity of the Earth's climate system, which is certainly not the case for simple climate models.

Since the TAR, EMICs have been intensively 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 organised model intercomparisons.

Figure 8.8.1 (adapted from Petoukhov et al., 2005) compares the results for present-day climate of some of the EMICs utilised in Chapter 10 for long-term climate change projections (see Table 8.8.2) with observational data and results of GCMs that took part in the AMIP (Atmospheric Model Intercomparison Project) and CMIP1 (Coupled Model Intercomparison Project, phase 1) (Gates et al., 1999; Lambert and Boer, 2001). From Figures 8.8.1a and 8.8.1b, it can be seen that the simulated latitudinal distributions of the zonally averaged surface air temperature for boreal winter and boreal summer are in rather 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 8.8.1c and 8.8.1d indicate that EMICs satisfactorily reproduce the general structure of the observed zonally averaged precipitation. Here again, for most latitudes, the results of EMICs are within the range of GCM results. 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 (Figure 8.8.2). This increase falls by and large within the range of GCM results, though on average, EMICs tend to yield slightly smaller temperature changes, i.e., EMICs seem to have, on average, a slightly weaker climate sensitivity than GCMs.

[INSERT FIGURE 8.8.1 HERE]

[INSERT FIGURE 8.8.2 HERE]

The responses of the North Atlantic meridional overturning circulation to increasing atmospheric CO_2 concentration and idealised freshwater perturbations as simulated by EMICs have also been compared to those obtained by coupled GCMs (Gregory et al., 2005; Stouffer et al., 2005). 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 experience a characteristic hysteresis response of the North Atlantic meridional overturning circulation 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 the one discussed in Brovkin et al. (2005). In this study, 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 a deforestation of about 15×10^6 km^2, all models exhibit a decrease in globally averaged, annual mean surface temperature in the range of 0.13–0.25°C. Further experiments in which historical changes in atmospheric CO_2 concentration were prescribed reveal that, for the whole last millennium, the biogeophysical cooling due to deforestation is less pronounced than the warming induced by increased atmospheric CO_2 level (0.27–0.62°C). During the 19th century, the cooling effect due to deforestation appears to counterbalance, albeit not completely, the warming effect of increasing CO_2 concentration.

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Tables

Table 8.2.1. Selected Model Features: Salient features of the participating AR4 coupled models 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 horizontal and vertical resolution of the model atmosphere and ocean, 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 AR4 coupled 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
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
BCC-CM1, ?	Beijing Climate Center, China	NA	NA	NA	NA	NA
CCSM3, 2005	National Center for Atmospheric Research, USA	top = 2.2 hPa T85(1.4° × 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	layers, canopy, routing Oleson et al., 2004; Branstetter, 2001
CGCM3.1(T47), 2005	Canadian Centre for Climate Modeling & Analysis, Canada	top = 1 hPa T47(~2.8° × 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
CGCM3.1(T63), 2005	Canadian Centre for Climate Modeling & Analysis, Canada	top = 1 hPa T63(~1.9° × 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
CNRM-CM3, 2004	Météo-France/Centre National de Recherches Météorologiques, France	top = 0.05 hPa T63(~1.9° × 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
CSIRO-MK3.0, 2001	CSIRO Atmospheric Research, Australia	top = 4.5 hPa T63(~1.9° × 1.9°)L18 Gordon et al., 2002	0.8° × 1.9° L31 depth, rigid lid Packanowski, 1996	rheology, leads O'Farrell, 1998; Semtner, 1976	no adjustments Gordon et al., 2002	layers, canopy Gordon et al., 2002

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
ECHAM5/MPI-OM, 2005	Max Planck Institute for Meteorology, Germany	top = 10 hPa T63(~1.9° × 1.9°)L31 Roeckner et al., 2003	1.5° × 1.5° L40 depth, free surface Marsland et al., 2003	rheology, leads Hibler, 1979; Semtner, 1976	no adjustments	bucket, canopy, routing Hagemann, 2002; Hagemann & Dümenil–Gates, 2001
ECHO-G, 1999	Meteorological Institute of the University of Bonn, Meteorological Research Institute of KMA, and Model & Data Group, Germany/Korea	top = 10 hPa T30(~3.9° × 3.9°)L19 Roeckner et al., 1996	0.5–2.8° × 2.8°L20 depth, free surface Wolff et al., 1997	rheology, leads Wolff et al., 1997	heat, freshwater Min et al., 2004, 2005	bucket, canopy, routing Roeckner et al., 1996; Dümenil-Todini, 1992
FGOALS-g1.0, 2004	LASG/Institute of Atmospheric Physics, China	top = 2.2 hPa T42(~2.8° × 2.8°)L26 Wang et al., 2004	1.0° × 1.0° L16 eta, free surface Zhang et al., 2003	rheology, leads Liu et al., 2004	no adjustments Yu et al. 2002, 2004	layers, canopy,routing Bonan et al., 2002; Branstetter, 2001
GDFL-CM2.0, 2005	U.S. Dept. of Commerce/NOAA/ Geophysical Fluid Dynamics Laboratory, USA	top = 3 hPa 2.0° × 2.5° L24 GFDL GAMDT, 2004	0.3–1.0° × 1.0° depth, free surface Gnanadesikan et al., 2004	rheology, leads? Winton, 2000; Delworth et al., 2004	no adjustments Delworth et al., 2004	bucket, canopy, routing Milly-Shmakin, 2002; GFDL GAMDT, 2004
GDFL-CM2.1, 2005	U.S. Dept. of Commerce/NOAA/ Geophysical Fluid Dynamics Laboratory, USA	top = 3 hPa 2.0° × 2.5° L24 GFDL GAMDT, 2004 with semi-Lagrangian transports	0.3–1.0° × 1.0° depth, free surface Gnanadesikan et al., 2004	rheology, leads? Winton, 2000; Delworth et al., 2004	no adjustments Delworth et al., 2004	bucket, canopy, routing Milly-Shmakin, 2002; GFDL GAMDT, 2004
GISS-AOM, 2004	NASA/Goddard Institute for Space Studies, USA	top = ? 3° × 4° L12 Russell et al., 1995	3 × 4° L16 mass/area, free sfc. Russell et al., 1995	rheology, leads Russell et al., 1995	no adjustments Russell et al., 1995	layers, canopy, routing Hansen et al., 1983
GISS-EH, 2004	NASA/Goddard Institute for Space Studies, USA	top = ? 4° × 5° L12 Schmidt et al., 2005	4° × 5° L16 density, free surface Bleck, 2002	rheology, leads Schmidt et al., 2004	no adjustments	layers, canopy, routing Friend-Kiang, 2005
GISS-ER, 2004	NASA/Goddard Institute for Space Studies, USA	top = ? 4° × 5° L12 Schmidt et al., 2005	3° × 4° L16 mass/area, free sfc. Russell et al., 1995	rheology, leads Liu et al., 2003	no adjustments	layers, canopy, routing Friend-Kiang, 2005

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
INM-CM3.0, 2004	Institute for Numerical Mathematics, Russia	top = 10 hPa 4º × 5º L21 Alekseev et al., 1998; Galin et al., 2003	2º × 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
IPSL-CM4,	Institut Pierre Simon Laplace, France	top = ? hPa 2.5º × 3.75º L19 Marti et al., 2005	1–2º × 2º L? depth, free surface Madec et al., 1998	rheology, leads Fichefet et al., 1997; Gosse-Fichefet, 1999	no adjustments Marti et al., 2005	layers, canopy, routing Krinner et al., 2005
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(~1.1º × 1.1º)L56 K-1 Developers, 2004	0.2º × 0.3ºL47 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
MIROC3.2(medres), 2004	Center for Climate System Research (University of Tokyo), National Institute for Environmental Studies, and Frontier Research Center for Global Change (JAMSTEC), Japan	top = 30 km T42(~2.8º × 2.8º)L20 K-1 Developers, 2004	0.5–1.4º × 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
MRI-CGCM2.3.2, 2003	Meteorological Research Institute, Japan	top = 0.4 hPa T42(~2.8º × 2.8º)L30 Shibata et al., 1999	0.5–2.0º × 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
PCM, 1998	National Center for Atmospheric Research, USA	top = 2.2 hPa T42(~2.8º × 2.8º)L26 Kiehl et al., 1998	0.5–0.7º × 1.1º L40 depth, free surface Maltrud et al., 1998	rheology, leads Hunke-Ducowicz, 1997, 2003; Zhang et al., 1999	no adjustments Washington et al., 2000	layers, canopy, no routing Bonan, 1998
UKMO-HadCM3, 1997	Hadley Centre for Climate Prediction and Research/Met Office, UK	top = 5 hPa 2.5º × 3.8º L19 Pope et al., 2000	1.5º × 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
UKMO-HadGEM, 2004	Hadley Centre for Climate Prediction and Research/Met Office, UK	top = 39.2 km ~1.3º × 1.9º L38 Martin et al., 2004	0.3–1.0º × 1.0º L40 depth, free surface Roberts, 2004	rheology, leads Hunke-Dukowicz, 1997; Semtner, 1976; Lipscomb, 2001	no adjustments Johns et al., 2004	layers, canopy, routing Essery et al., 2001; Oki-Sud, 1998

Table 8.8.1. Simple climate model parameter values utilised to simulate coupled GCM results from the IPCC AR4 dataset. Other parameters are as used in the TAR (Table 9.A1).

AOGCM	F_{2x} (W m^{-2})	T_{2x} (^oC)	ΔT^+ (^oC)	k (cm^2 s^{-1})	RLO	LO and NS (W m^{-2} $^oC^{-1}$)
CNRM_CM3	3.71[a]	2.05	10.0[b]	1.98	1.24	0.5
GFDL_CM2.0	3.71[a]	1.85	10.0[b]	2.33	1.48	0.5
GISS-EH	3.71[a]	2.86	8.7	10.61	1.55	0.5
MIROC3.2(medres)	3.66	3.81	7.2	3.84	1.37	0.5

Notes:
F_{2x}: radiative forcing for a doubled CO_2 concentration.
T_{2x}: climate sensitivity.
ΔT^+: magnitude of warming that would result in a collapse of the THC.
k: ocean effective vertical diffusivity.
RLO: ratio of the equilibrium temperature changes over land versus ocean.
LO and NS: land-ocean and Northern Hemisphere-Southern Hemisphere exchange coefficients.
(a) The best estimate from Myhre et al. (1998) is used.
(b) Default value.

Table 8.8.2. 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	LAND SURFACE	BIOSPHERE	INLAND ICE
BERN2.5D (Plattner et al., 2002)	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)	NST, NSM (Schmittner and Stocker, 1999)	BO (Marchal et al., 1998), BT (Siegenthaler and Oeschger, 1987)	
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)	NST, NSM, RIV (Edwards and Marsh, 2005)		
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 (Pethoukhov et al., 2000)	1-LST, CSM, RIV (Pethoukhov 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)
CLIMBER-3α	SD, 3-D, CRAD, ICL, 7.5° × 22.5°, L10 (Pethoukhov et al., 2000)	PE, 3-D, FS, ISO, MESO, TCS, DC*, 3.75° × 3.75°, L24	M-LT, R, 2-LIT (Fichefet and Morales Maqueda, 1997)	1-LST, CSM, RIV (Pethoukhov et al., 2000)	BO* (Six and Maier-Reimer, 1996), BT* (Brovkin et al., 2002), BV* (Brovkin et al., 2002)	
LOVECLIM (Renssen et al., 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)	M-LT, R, 2-LIT (Fichefet and Morales Maqueda, 1997)	1-LST, BSM, RIV (Opsteegh et al., 1998)	BO (Mouchet and François, 1997), BT (Brovkin et al., 2002), BV (Brovkin et a., 2002)	TM, 3-D, 10 km × 10 km, L30 (Huybrechts, 2002)
MIT-IGSM2 (Sokolov et al., 2005)	SD, 2-D(φ, z), CRAD, ICL, CHEM*, 4°, L11 (Sokolov and Stone, 1998)	PE, 3-D, FS, ISO, MESO, 4° × 4°, L15 (Marshall et al., 1997)	M-LT, 2-LIT (Winton, 2000)	M-LST, CSM (Bonan et al., 2002)	BO (McKinley et al., 2004), BT, BV*	
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)	1-LST, BSM (Gallée et al., 1991)	BO* BT* (Brovkin et al., 2002), BV (Brovkin et al., 2002)	M, 1-D(φ), 0.5° (Crucifix and Berger, 2002)
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)	M-LT, R, M-LIT (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-dimesnional; LRAD = linearised radiation scheme; CRAD = comprehensive radiation scheme; NCL = non-interactive cloudiness; ICL = interactive cloudiness; CHEM = interactive chemistry; 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 "Lmm", where mm 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 "Lmm", where mm is the number of vertical levels.
Sea ice: 0-LT = zero-layer thermodynamic scheme; M-LT = multi-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); M-LIT = multi-level ice thickness distribution.
Land surface: NST = no explicit computation of soil temperature; 1-LST = one-layer soil temperature scheme ; M-LST = multi-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 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 carbon dynamics; BV = dynamical vegetation model.
Inland ice: TM = thermomechanical model M = mechanical model (isothermal); × kilometres; the vertical resolution is expressed as "Lmm", where mm 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