Chapter 9: Understanding and Attributing Climate Change

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Supplementary Material

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

Executive Summary

Evidence of the effect of external influences on the climate system has continued to accumulate since the TAR. The evidence now available is substantially stronger and is based on analyses of widespread temperature increases throughout the climate system and changes in other climate variables.

Human-induced warming of the climate system is widespread. Anthropogenic warming of the climate system can be detected in temperature observations taken at the surface, in the troposphere and in the oceans. Multi-signal detection and attribution analyses, which quantify the contributions of different natural and anthropogenic forcings to observed changes, show that greenhouse gas forcing alone during the past half century would likely have resulted in greater than the observed warming if there had not been an offsetting cooling effect from aerosol and other forcings.

It is extremely unlikely (<5%) that the global pattern of warming during the past half century can be explained without external forcing, and very unlikely that it is due to known natural external causes alone. The warming occurred in both the ocean and the atmosphere and took place at a time when natural external forcing factors would likely have produced cooling.

Greenhouse gas forcing has very likely caused most of the observed global warming over the last 50 years. This conclusion takes into account observational and forcing uncertainty, and the possibility that the response to solar forcing could be underestimated by climate models. It is also robust to the use of different climate models, different methods for estimating the responses to external forcing, and variations in the analysis technique.

Further evidence has accumulated of an anthropogenic influence on the temperature of the free atmosphere as measured by radiosondes and satellite-based instruments. The observed pattern of tropospheric warming and stratospheric cooling is very likely due to the influence of anthropogenic forcing, particularly greenhouse gases and stratospheric ozone depletion. The combination of a warming troposphere and a cooling stratosphere has likely led to an increase in the height of the tropopause. It is likely that anthropogenic forcing has contributed to the general warming observed in the upper several hundred meters of the ocean during the latter half of the 20th century. Anthropogenic forcing, resulting in thermal expansion from ocean warming and glacier mass loss, has very likely contributed to sea level rise during the latter half 20th century. It is difficult to quantify the contribution of anthropogenic forcing to ocean heat content increase and glacier melting with presently available detection and attribution studies.

It is likely that there has been a substantial anthropogenic contribution to surface temperature increases in every continent except Antarctica since the middle of the 20th century. Anthropogenic influence has been detected in every continent except Antarctica (which has insufficient observational coverage to make an assessment), and in some sub-continental land areas. The ability of coupled climate models to simulate the temperature evolution on continental scales and the detection of anthropogenic effects on each of six continents provides stronger evidence of human influence on the global climate than was available to the TAR. No climate model that has used natural forcing only has reproduced the observed global mean warming trend or the continental mean warming trends in all individual continents (except Antarctica) over the second half of the 20th century.

Difficulties remain in attributing temperature changes on smaller than continental scales and over timescales of less than 50 years. Attribution at these scales has, with limited exceptions, not yet been established. Averaging over smaller regions reduces the natural variability less than does averaging over large regions, making it more difficult to distinguish between changes expected from different external forcings, or between external forcing and variability. Also, temperature changes associated with some modes of variability are poorly simulated by models in some regions and seasons. Furthermore, the small-scale details of external forcing and the response simulated by models are less credible than large-scale features.

Surface temperature extremes have likely been affected by anthropogenic forcing. Many indicators of climate extremes and variability, including the annual numbers of frost days, warm and cold days, and warm and cold nights, show changes that are consistent with warming. An anthropogenic influence has been detected in some of these indices, and there is evidence that anthropogenic forcing may have substantially increased the risk of extremely warm summer conditions regionally, such as the 2003 European heat wave.

There is evidence of anthropogenic influence in other parts of the climate system. Anthropogenic forcing has likely contributed to recent decreases in Arctic sea ice extent and to glacier retreat. The observed decrease in global snow cover extent and the widespread retreat of glaciers are consistent with warming, and there is evidence that this melting has likely contributed to sea level rise.

Trends over recent decades in the Northern and Southern Annular Modes, which correspond to sea level pressure reductions over the poles, are likely related in part to human activity, affecting storm tracks, winds, and temperature patterns in both hemispheres. Models reproduce the sign of the Northern Annular Mode trend, but the simulated response is smaller than observed. Models including both greenhouse gas and stratospheric ozone changes simulate a realistic trend in the Southern Annular Mode, leading to a detectable human influence on global sea level pressure patterns.

The response to volcanic forcing simulated by some models is detectable in global annual mean land precipitation during the latter half of the 20th century. The latitudinal pattern of change in land precipitation and observed increases in heavy precipitation over the 20th century appear to be consistent with the anticipated response to anthropogenic forcing. It is more likely than not that anthropogenic influence has contributed to increases in the frequency of the most intense tropical cyclones. Stronger attribution to anthropogenic factors is not possible at present because the observed increase in the proportion of such storms appears to be larger than suggested by either theoretical or modelling studies and because of inadequate process knowledge, insufficient understanding of natural variability, uncertainty in modelling intense cyclones and uncertainties in historical tropical cyclone data.

Analyses of paleo-climate data have increased confidence in the role of external influences on climate. Coupled climate models used to predict future climate have been used to understand past climatic conditions of the Last Glacial Maximum and the Mid-Holocene. While many aspects of these past climates are still uncertain, key features have been reproduced by climate models using boundary conditions and radiative forcing for those periods. A substantial fraction of the reconstructed Northern Hemisphere interdecadal temperature variability of the seven centuries prior to 1950 is very likely attributable to natural external forcing, and it is likely that anthropogenic forcing contributed to the early 20th century warming evident in these records.

Estimates of the climate sensitivity are now better constrained by observations. Estimates based on observational constraints indicate that it is very likely that the equilibrium climate sensitivity is larger than 1.5°C with a most likely value between 2°C and 3°C. The upper 95% limit remains difficult to constrain from observations. This supports the overall assessment based on modelling and observational studies that the equilibrium climate sensitivity is likely 2 to 4.5°C with a most likely value of approximately 3°C. (Chapter 10, Box 10.2). The transient climate response, based on observational constraints, is very likely larger than 1°C and very unlikely to be greater than 3.5°C at the time of CO_2 doubling in response to a 1% per year increase in CO_2, supporting the overall assessment that the transient climate response is very unlikely greater than 3°C (Chapter 10).

Overall consistency of evidence. Many observed changes in surface and free atmospheric temperature, ocean temperature and sea-ice extent, and some large-scale changes in the atmospheric circulation over the 20th century are distinct from internal variability and consistent with the expected response to anthropogenic forcing. The simultaneous increase in energy content of all the major components of the climate system as well as the magnitudes and patterns of warming within and across the different components supports the conclusion that the cause of the warming is extremely unlikely (<5%) to be the result of internal processes. Qualitative consistency is also apparent in some other observations, including snow cover, glacier retreat, and heavy precipitation.

Remaining uncertainties. Further improvements in models and analysis techniques have led to increased confidence in the understanding of the influence of external forcing on climate since the TAR. However, estimates of some radiative forcings remain uncertain, including aerosol forcing and interdecadal variations in solar forcing. The net aerosol forcing over the 20th century from inverse estimates based on the observed warming likely ranges between –1.7 and –0.1 W m^{-2}. The consistency of this result with forward estimates of total aerosol forcing (Chapter 2) strengthens confidence in estimates of total aerosol forcing, despite remaining uncertainties. Nevertheless, the robustness of surface temperature attribution results to forcing and response uncertainty has been evaluated with a range of models, forcing representations and analysis procedures. The potential impact of the remaining uncertainties has been considered, to the extent possible, in the overall assessment of every line of evidence listed above. There is less confidence in the understanding of forced changes in other variables, such as surface pressure and precipitation, and on smaller spatial scales.

Better understanding of instrumental and proxy-climate records, and climate model improvements have increased confidence in climate model simulated internal variability. However, uncertainties remain. For example, there are apparent discrepancies between estimates of ocean heat content variability from models and observations. While reduced relative to the situation at the time of the TAR, uncertainties in the radiosonde and satellite records still affect confidence in estimates of the anthropogenic contribution to tropospheric temperature change. Incomplete global data sets and remaining model uncertainties still restrict our understanding of changes in extremes and attribution of changes to causes, although understanding of changes in the intensity, frequency and risk of extremes has improved.

9.1 Introduction

The objective of this chapter is to assess scientific understanding about the extent to which the observed climate changes that are reported in Chapters 3 to 6 are expressions of natural internal climate variability and/or externally forced climate change. The scope of this chapter includes "detection and attribution" but is wider than that of previous detection and attribution chapters in the SAR (Santer et al., 1996a) and the TAR (Mitchell et al., 2001). Climate models, physical understanding of the climate system, and statistical tools, including formal climate change detection and attribution tools, are used to interpret observed changes where possible. The detection and attribution research discussed in this chapter includes research on regional scales, extremes and variables other than temperature. This new work is placed in the context of a broader understanding of a changing climate. However, the ability to interpret some changes, particularly for non-temperature variables, is limited by uncertainties in the observations, physical understanding of the climate system, climate models, and external forcing estimates. Research on the impacts of these observed climate changes is assessed in Working Group II, Chapter 1.

9.1.1 What is Climate Change and Climate Variability?

Climate change "refers to a change in the state of the climate that can be identified (e.g., using statistical tests) by changes in the mean and/or the variability of its properties, and that persists for an extended period, typically decades or longer" (see Glossary). Climate change may be due to internal processes and/or external forcings. Some external influences, such as changes in solar radiation and volcanism, occur naturally and contribute to the total natural variability of the climate system. Other external changes, such as the change in composition of the atmosphere that began with the industrial revolution, are the result of human activity. A key objective of this chapter is to understand climate changes that result from anthropogenic and natural external forcings, and how they may be distinguished from changes and variability that result from internal climate system processes.

Internal variability is present on all time scales. Atmospheric processes that generate internal variability are known to operate on timescales ranging from virtually instantaneous (e.g., condensation of water vapour in clouds) up to years (e.g., troposphere-stratosphere or inter-hemispheric exchange). Other components of the climate system, such as the ocean and the large ice sheets tend to operate on longer time scales. These components produce internal variability of their own accord and also integrate variability from the rapidly varying atmosphere (Hasselmann, 1976). In addition, internal variability is also produced by coupled interactions between components, such as is the case with the El-Niño Southern Oscillation (ENSO; see Chapters 3 and 8).

Distinguishing between the effects of external influences and internal climate variability requires careful comparison between observed changes and those that are expected to result from external forcing. These expectations are based on physical understanding of the climate system. Physical understanding is based on physical principles. This understanding can take the form of conceptual models or it might be quantified with climate models that are driven with physically based forcing histories. An array of climate models is used to quantify expections in this way, ranging from simple energy balance models, to models of intermediate complexity, to comprehensive coupled climate models (Chapter 8) such as those that contributed to the multi-model data archive at PCMDI (MMD). The latter have been extensively evaluated by their developers and a broad investigator community. The extent to which a model is able to reproduce key features of the climate system and its variations, for example the seasonal cycle, increases its credibility for simulating changes in climate.

The comparison between observed changes and those that are expected is performed in a number of ways. Formal detection and attribution (Section 9.1.2) uses objective statistical tests to assess whether observations contain evidence of the expected responses to external forcing that is distinct from variation generated within the climate system (internal variability). These methods generally do not rely on simple linear trend analysis. Instead, they attempt to identify in observations the responses to one or several forcings by exploiting the time and/or spatial pattern of the expected responses. The response to forcing does not necessarily evolve over time as a linear trend, either because the forcing itself may not evolve in that way, or because the response to forcing is not necessarily linear.

The comparison between model simulated and observed changes, for example, in detection and attribution methods (Section 9.1.2), also carefully accounts for the effects of changes over time in the availability of climate observations to ensure that a detected change is not an artefact of a changing observing system. This is usually done by evaluating climate model data only where and when observations are available, in order to mimic the observational system and avoid possible biases introduced by changing observational coverage.

9.1.2 What is Climate Change Detection and Attribution?

The concepts of climate change *detection* and *attribution* used in this chapter remain as they were defined in the TAR (IPCC, 2001; Mitchell et al., 2001). *Detection* "is the process of demonstrating that climate has changed in some defined statistical sense, without providing a reason for that change" (Glossary). In this chapter, the methods used to identify change in observations are based on the expected responses to external forcing (Section 9.1.1), either from physical understanding or as simulated by climate models. An identified change is *detected* in observations if its likelihood of occurence by chance due to internal variability alone is determined to be small. A failure to detect a particular response might occur for a number of reasons, including the possibility that the response is weak relative to internal variability, or that the metric used to measure change is insensitive to the expected change. For example, the annual global mean precipitation may not be a sensitive indicator of the influence of increasing greenhouse concentrations given the expectation that greenhouse forcing would result in moistening in some latitudes that is partially offset by drying elsewhere (Chapter 10; see also Section 9.5.4.2). Furthermore, because detection studies are statistical in nature, there is always some small possibility of spurious detection. The risk of such a possibility is reduced when corroborating lines of evidence provide a physically consistent view of the likely cause for the detected changes and render then less consistent with internal variability (see, for example, Section 9.7).

Many studies use climate models to predict the expected responses to external forcing, and these predictions are usually represented as patterns of variation in space, time, or both (see Chapter 8 for an evaluation). Such patterns, or *fingerprints*, are usually derived from changes simulated by a climate model in response to forcing. Physical understanding can also be used to develop conceptual models of the anticipated pattern of response to external forcing and the consistency between responses in different variables and different parts of the climate system. For example, precipitation and temperature are ordinarily inversely correlated in some regions, with increases in temperature corresponding to drying conditions. Thus a warming trend in such a region that is not associated with rainfall change may indicate an external influence on the climate of that region (Nicholls et al., 2005) (Section 9.4.2.3). Purely diagnostic approaches can also be used. For example, Schneider and Held (2001) use a technique that discriminates between slow changes in climate and shorter time scale variability to identify in observations a pattern of surface temperature change that is consistent with the expected pattern of change from anthropogenic forcing.

The spatial and temporal scales used to analyze climate change are carefully chosen so as to focus on the spatio-temporal scale of the response, filter out as much internal variability as possible (often by using a metric that reduces the influence of internal variability, see Appendix 9.A) and enable the separation of the responses to different forcings. For example, it is expected that greenhouse gas forcing would cause a large-scale pattern of warming that evolves slowly over time, and thus analysts often smooth data to remove small-scale variations. Similarly, when fingerprints from AOGCMs are used, averaging over an ensemble of coupled model simulations helps separate the model's response to forcing from its simulated internal variability.

Detection does not imply attribution of the detected change to the assumed cause. *Attribution* "of causes of climate change is the process of establishing the most likely causes for the detected change with some defined level of confidence" (see Glossary). As noted in the SAR (IPCC, 1996) and the TAR (IPCC, 2001), unequivocal attribution would require controlled experimentation with our climate system. Since that is not possible, in practice attribution of anthropogenic climate change is understood to mean demonstration that a detected change is "consistent with the estimated responses to the given combination of anthropogenic and natural forcing" and "not consistent with alternative, physically-plausible explanations of recent climate change that exclude important elements of the given combination of forcings" (IPCC, 2001).

The consistency between an observed change and the estimated response to a hypothesized forcing is often determined by estimating the amplitude of the hypothesized pattern of change from observations and then assessing whether this estimate is statistically consistent with the expected amplitude of the pattern. Attribution studies additionally assess whether the response to a key forcing, such as greenhouse gas increases, is distinguishable from that to other forcings (Appendix 9.A). These questions are typically investigated using a multiple regression of observations onto several fingerprints representing climate responses to different forcings that, ideally, are clearly distinct from each other (i.e., as distinct spatial patterns or distinct evolutions over time; see Section 9.2.2). If the response to this key forcing can be distinguished, and if even rescaled combinations of the responses to other forcings do not sufficiently explain the observed climate change, then the evidence for a causal connection is substantially increased. For example, the attribution of recent warming to greenhouse gas forcing becomes more reliable if the influences of other external forcings, for example solar forcing, are explicitly accounted for in the analysis. This is an area of research with considerable challenges because different forcing factors may lead to similar large-scale spatial patterns of response (Section 9.2.2). Note that another key element in attribution studies is the consideration of the physical consistency of multiple lines of evidence.

Both detection and attribution require knowledge of the internal climate variability on the timescales considered, usually decades or longer. The residual variability that remains in instrumental observations after the estimated effects of external forcing have been removed is sometimes used to estimate internal variability. However, these estimates are uncertain because the instrumental record is too short to give a well-constrained estimate of internal variability, and because of uncertainties in the forcings and the estimated responses. Thus internal climate variability is usually estimated from long control simulations from coupled climate models. Subsequently, an assessment is usually made of the consistency between the residual variability referred to above and the model based estimates of internal variability; analyses that yield unplausibly large residuals are not considered credible (for example, this might happen if an important forcing is missing, or if the internal variability from the model is too small). Confidence is further increased by systematic intercomparison of the ability of models to simulate the various modes of observed variability (Chapter 8), by comparisons between variability in observations and climate model data (Section 9.4), and by comparisons between proxy reconstructions and climate simulations of the last millennium (Chapter 6 and Section 9.3).

Studies where the estimated pattern amplitude is substantially different from that simulated by models can still provide some understanding of climate change but need to be treated with caution (examples are given in Section 9.5). If this occurs for variables where confidence in the climate models is limited, such a result may simply reflect weaknesses in models. On the other hand, if this occurs for variables where confidence in the models is higher, it may raise questions about the forcings, such as whether all important forcings have been included or whether they have the correct amplitude, or questions about uncertainty in the observations.

Model and forcing uncertainties are important considerations in attribution research. Ideally, the assessment of model uncertainty should include uncertainties in model parameters (for example, as explored by multi-model ensembles), and in the representation of physical processes in models (structural uncertainty). Such a complete assessment is not yet available, although model intercomparison studies (Chapter 8) improve the understanding of these uncertainties. The effects of forcing uncertainties, which can be considerable for some forcing agents such as solar and aerosol forcing (Section 9.2), also remain difficult to evaluate despite advances in research. Detection and attribution results based on several models or several forcing histories do provide information on the effects of model and forcing uncertainty. Such studies suggest that while model uncertainty is important, key results, such as attribution of a human influence on temperature change during the latter half of the 20th century, are robust.

Detection of anthropogenic influence is not yet possible for all climate variables for a variety of reasons. Some variables respond less strongly to external forcing, or are less reliably modelled or observed. In these cases, research that describes observed changes and offers physical explanations, for example, by demonstrating links to sea surface temperature changes, contributes substantially to the understanding of climate change and is therefore discussed in this chapter.

The approaches used in detection and attribution research described above cannot fully account for all uncertainties, and thus ultimately expert judgement is required to give a calibrated assessment of whether a specific cause is responsible for a given climate change. The assessment approach used in this chapter is to consider results from multiple studies using a variety of observational data sets, models, forcings, and analysis techniques. The assessment based on these results typically takes into account the number of studies, the extent to which there is consensus amongst studies on the significance of detection results, the extent to which there is consensus on the consistency between the observed change and the change expected from forcing, the degree of consistency with other types of evidence, the extent to which known uncertainties are accounted for in and between studies, and whether there might be other physically-plausible explanations for the given climate change. Having determined a particular likelihood assessment, this was then further downweighted to take into account any remaining uncertainties, such as, for example, structural uncertainties or a limited exploration of possible forcing histories of uncertain forcings. The overall assessment also considers whether several independent lines of evidence strengthen a result.

While the approach used in most detection studies assessed in this chapter is to determine whether observations exhibit the expected response to external forcing, for many decision-makers a question posed in a different way may be more relevant. For instance, they may ask, “Are the continuing drier-than-normal conditions in the Sahel due to human causes?” Such questions are difficult to respond to because of a statistical phenomenon known as “selection bias”. The fact that the questions are “self selected” from the observations (only large observed climate anomalies in a historical context would be likely to be the subject of such a question) makes it difficult to assess their statistical significance from the same observations (see for example von Storch and Zwiers, 1999). Nevertheless, there is a need for answers to such questions, and examples of studies that attempt to do so are discussed in this chapter (e.g., see Section 9.4.3.3).

9.1.3 The Basis from which we Begin

Evidence of a human influence on the recent evolution of the climate has accumulated steadily during the past two decades. The first IPCC Assessment Report (IPCC, 1990) contained little observational evidence of a detectable anthropogenic influence on climate. However, six years later the IPCC WG1 Second Assessment Report (IPCC, 1996) concluded that “the balance of evidence” suggested there had been a “discernible” human influence on the climate of the 20th century. Considerably more evidence accumulated during the subsequent five years, such that the TAR (IPCC, 2001) was able to draw a much stronger conclusion, not just on the detectability of a human influence, but on its contribution to climate change during the 20th century.

The evidence that was available at the time of the TAR was considerable. Using results from a range of detection studies of the instrumental record, which was assessed using fingerprints and estimates of internal climate variability from several climate models, it was found that the warming over the 20th century was “very unlikely to be due to internal variability alone as estimated by current models”.

Simulations of global mean 20th century temperature change that accounted for anthropogenic greenhouse gases and sulphate aerosols as well as solar and volcanic forcing were found to be generally consistent with observations. In contrast, a limited number of simulations of the response to known natural forcings alone indicated that these may have contributed to the observed warming in the first half of the 20th century, but could not provide an adequate explanation of the warming in the second half of the 20th century, nor the observed changes in the vertical structure of the atmosphere.

Attribution studies had begun to use techniques to determine whether there was evidence that the responses to several different forcing agents were simultaneously present in observations, mainly of surface temperature and of temperature in the free atmosphere. A distinct greenhouse gas signal was found to be detectable whether or not other external influences were explicitly considered and the amplitude of the simulated greenhouse gas response was generally found to be consistent with observationally based estimates on the scales that were considered. Also, in most studies, the estimated rate and magnitude of warming over the second half of the 20th century due to increasing greenhouse gas concentrations alone was comparable with, or larger than, the observed warming. This result was found to be robust to attempts to account for uncertainties, such as observational uncertainty and sampling error in estimates of the response to external forcing, as well as differences in assumptions and analysis techniques.

The TAR also reported on a range of evidence of qualitative consistencies between observed climate changes and model responses to anthropogenic forcing, including global temperature rise, increasing land-ocean temperature contrast, diminishing Arctic sea ice extent, glacial retreat and increases in precipitation at high Northern latitudes.

A number of uncertainties remained at the time of the TAR. For example, large uncertainties remained in estimates of internal climate variability. However, even substantially inflated (doubled or more) estimates of model simulated internal variance were found unlikely to be large enough to nullify the detection of an anthropogenic influence on climate. Uncertainties in external forcing were also reported, particularly in anthropogenic aerosol, solar and volcanic forcing, and in the magnitude of the corresponding climate responses. These uncertainties contributed to uncertainties in detection and attribution studies. Particularly, estimates of the contribution to the 20th century warming by natural forcings and anthropogenic forcings other than greenhouse gases showed some discrepancies with climate simulations and were model dependent. These results made it difficult to attribute the observed climate change to one specific combination of external influences.

Based on the available studies and understanding of the uncertainties, the TAR concluded that "in the light of new evidence and taking into account the remaining uncertainties, most of the observed warming over the last 50 years is likely to have been due to the increase in greenhouse gas concentrations". Since the TAR, a larger number of model simulations using more complete forcings have become available, evidence on a wider range of variables has been analyzed, and many important uncertainties have been further explored, and in many cases reduced. These advances are assessed in this chapter.

9.2 Radiative Forcing and Climate Response

This section briefly summarizes the understanding of radiative forcing based on the assessment in Chapter 2, and of the climate response to forcing. Uncertainties in the forcing and estimates of climate response, and their implications for understanding and attributing climate change are also discussed. The discussion of radiative forcing focuses primarily on the period since 1750, with a brief reference to periods in the more distant past that are also assessed in the chapter, such as the last millennium, the Last Glacial Maximum and the Mid-Holocene.

Two basic types of calculations have been used in detection and attribution studies. The first uses best estimates of forcing together with best estimates of modelled climate processes to calculate the effects of external changes in the climate system (forcings) on the climate (the response). These "forward calculations" can then be directly compared to the observed changes in the climate system. Uncertainties in these simulations result from uncertainties in the radiative forcings that are used, and from model uncertainties that affect the simulated response to the forcings. Forward calculations are explored in this chapter and compared to observed climate change.

Results from forward calculations are used for formal detection and attribution analyses. In such studies a climate model is used to calculate response patterns ("fingerprints") for individual forcings or sets of forcings which are then combined linearly to provide the best fit to the observations. This procedure assumes that the amplitude of the large-scale pattern of response scales linearly with the forcing, and that patterns from different forcings can be added to obtain the total response. This assumption may not hold for every forcing, particularly not on smaller spatial scales and may be violated when forcings interact non-linearly (for example, black carbon absorption decreases cloudiness and thereby decreases the indirect effects of SO4 aerosol). Generally, however, the assumption is expected to hold for most forcings (e.g., Penner et al., 1997; Meehl et al., 2004). Errors or uncertainties in the magnitude of the forcing or the magnitude of a model's response to the forcing should not affect detection results provided that the space-time pattern of the response is correct. However, for the linear combination of responses to be considered consistent with the observations, the scalings on individual response patterns should indicate that the model does not need to be rescaled to match the observations (Sections 9.1.2, 9.4.1.4, and Appendix 9.A) given uncertainty in the amplitude of forcing, model response and estimate due to internal climate variability. For detection studies, if the space-time pattern of response is incorrect, then the scaling, and hence detection and attribution results, will be affected.

In the second type of calculation, the so-called "inverse" calculations, the magnitude of uncertain parameters in the forward model (including the forcing that is applied) is varied in order to provide a best fit to the observational record. In general, the greater the degree of a priori uncertainty in the parameters of the model, the more the model is allowed to adjust. Probabilistic posterior estimates for model parameters and uncertain forcings are obtained by comparing the agreement between simulations and observations, and taking into account prior uncertainties (including those in observations; see Sections 9.2.1.2, 9.6 and Appendix 9.B).

9.2.1 Radiative Forcing Estimates Used to Simulate Climate Change

9.2.1.1 Summary of "Forward" Estimates of Forcing for the Instrumental Period

Estimates of the radiative forcing (see Section 2.2 for a definition) since 1750 from forward model calculations and observations are reviewed in detail in Chapter 2 and provided in Table 2.12. Chapter 2 describes estimated forcing resulting from increases in long-lived greenhouse gases (CO_2, CH_4, N_2O, halocarbons), decreases in stratospheric ozone, increases in tropospheric ozone, sulphate aerosols, nitrate aerosols, black carbon (BC) and organic matter from fossil fuel burning, biomass burning aerosols, mineral dust aerosols, land use change, indirect aerosol effects on clouds, aircraft cloud effects, solar variability, and stratospheric and tropospheric H_2O increases from CH_4 and irrigation. An example of one model's implemented set of forcings is given in Figure 2.23. While some members of the multi-model dataset at PCMDI (MMD) have included a nearly complete list of these forcings for the purpose of simulating the 20th century climate (see Supplementary material, Table S9.1), most detection studies to date have used model runs with a more limited set of forcings. The combined anthropogenic forcing from the estimates in Section 2.9.2 since 1750 is 1.6 W m^{-2}, with a 90% range of 0.6 to 2.4 W m^{-2}, indicating that it is extremely likely that humans have exerted a substantial warming influence on climate over that time period. The forcing by greenhouse gas plus O3 combined is 2.9 ± 0.3 W m^{-2} and the total aerosol forcing (combined direct and indirect "cloud albedo" effect) is virtually certain to be negative and estimated to be –1.3 (90% uncertainty range of –2.2 to –0.5 W m^{-2}). In contrast, the direct radiative forcing due to increases in solar irradiance is estimated to be +0.12 (90% range from 0.06 to 0.3 W m^{-2}). Also, Chapter 2 concludes that it is exceptionally unlikely that the combined natural (solar and volcanic) radiative forcing has had a warming influence comparable to that of the combined anthropogenic forcing over the period 1950-2005. As noted in Chapter 2, the estimated global average surface temperature response from these forcings may have a different proportionality constant to the forcing for different forcings since all forcings do not have the same "efficacy" (i.e., effectiveness at changing the surface temperature compared to carbon dioxide). Thus summing these forcings does not necessarily give an adequate estimate of the response in global average surface temperature.

9.2.1.2 Summary of "Inverse" Estimates of Net Aerosol Forcing

Forward model approaches to estimating aerosol forcing are based on estimates of emissions and models for aerosol physics and chemistry. They directly resolve the separate contributions by various aerosol components and forcing mechanisms. This must be borne in mind when comparing results to those from inverse calculations (see Section 9.6 and Appendix 9.B for details), which, for example, infer the net aerosol forcing required to match climate model simulations with observations. These methods can be applied using a global average forcing and response, or using the spatial and temporal patterns of the climate response in order to increase the ability to distinguish between responses to different external forcings. Inverse methods have been used to constrain one or several uncertain radiative forcings, for example by aerosols, as well as climate sensitivity (Section 9.6) and other uncertain climate parameters (see Wigley, 1989; Schlesinger and Ramankutty, 1992; Wigley et al., 1997; Andronova and Schlesinger, 2001; Forest et al., 2001; Forest et al., 2002; Harvey and Kaufmann, 2002; Knutti et al., 2002, 2003; Andronova et al., 2005; Forest et al., 2006; see Table 9.1; Stott et al., 2006c). The reliability of the spatial and temporal patterns used is discussed in Sections 9.2.2.1 and 9.2.2.2.

In the past forward calculations have been unable to rule out a total net negative radiative forcing over the 20th century (Boucher and Haywood, 2001), However, Section 2.9 updates the Boucher and Haywood analysis for current radiative forcing estimates and shows that a net negative forcing is exceptionally unlikely (<1% probability). A net forcing close to zero would imply a very high value of climate sensitivity, and would be very difficult to reconcile with the observed increase in temperature (Sections 9.6, 9.7). Inverse calculations yield only the "net forcing", which includes all forcings that project on the fingerprint of the forcing that is estimated. For example, the response to tropospheric ozone forcing could project onto that for sulphate aerosol forcing. Therefore, differences between forward estimates and inverse estimates may have one of several causes including (1) the magnitude of the forward model calculation is incorrect due to inadequate physics and/or chemistry, (2) the forward calculation has not evaluated all forcings and feedbacks, or (3) other forcings project on the fingerprint of the forcing that is estimated in the inverse calculation.

Studies providing inverse estimates of aerosol forcing are compared in Table 9.1. One type of inverse method uses the ranges of climate change fingerprint scaling factors derived from detection and attribution analyses that attempt to separate the climate response to greenhouse gas forcing from that to aerosol forcing and, often, also natural forcing (Gregory et al., 2002a; Stott et al., 2006c, see also Section 9.4.1.4). These provide the range of fingerprint magnitudes (for example, for the combined temperature response to different aerosol forcings) that are consistent with observed climate change, and can therefore be used to infer the likely range of forcing that is consistent with the observed record. The separation between greenhouse gas and aerosol fingerprints exploits the fact that the forcing from well-mixed greenhouse gases is well known, and that errors in the model's transient sensitivity can therefore be separated from errors in aerosol forcing in the model (assuming that there are similar errors in a model's sensitivity to greenhouse gas and aerosol forcing; see Gregory et al., 2002a; Table 9.1). By scaling spatio-temporal patterns of response up or down, this technique takes account of gross model errors in climate sensitivity and net aerosol forcing but does not fully account for modelling uncertainty in the patterns of temperature response to uncertain forcings.

[INSERT TABLE 9.1 HERE]

Another approach uses the response of climate models, most often simple climate models or Earth System Models of Intermediate Complexity (EMICs, Chapter 8, Table 8.3**)** to explore the range of forcings and climate parameters that yield results consistent with observations (Andronova and Schlesinger, 2001; Forest et al., 2002; Harvey and Kaufmann, 2002; Knutti et al., 2002, 2003; Forest et al., 2006). Like detection methods, these approaches seek to fit the space-time patterns or spatial means in time, of observed surface, atmospheric or ocean temperatures. They determine the probability of combinations of climate sensitivity and net aerosol forcing based on the fit between simulations and observations (Section 9.6 and Appendix 9.B for further discussion). These are often based on Bayesian approaches, where prior assumptions for ranges of external forcing are used to constrain the estimated net aerosol forcing and climate sensitivity. Some of these studies use the difference between Northern and Southern Hemispheric mean temperature to separate the greenhouse gas and aerosol forcing effects (e.g., Andronova and Schlesinger, 2001; Harvey and Kaufmann, 2002). In these analyses, it is necessary to accurately account for hemispheric asymmetry in tropospheric ozone forcing in order to infer the hemiphseric aerosol forcing. Additionally, aerosols from biomass burning could cause an important fraction of the total aerosol forcing although this forcing shows little hemispheric asymmetry. Since it therefore projects on the GHG forcing, it is difficult to separate in an inverse calculation. Overall, results will be only as good as the spatial or time pattern that is assumed in the analysis. Missing forcings or lack of knowledge about uncertainties, and the highly parameterized spatial distribution of response in some of these models may hamper the interpretation of results.

Aerosol forcing appears to have grown rapidly during the period from 1945 to 1980, while greenhouse gas forcing grew more slowly (Ramaswamy et al., 2001). Global sulphur emissions (and thus, sulphate aerosol forcing) appear to have decreased after 1980 (Stern, 2005), further rendering the time evolution of aerosols and greenhouse gases distinct. As long as the temporal pattern of variation in aerosol forcing is approximately correct, the need to achieve a reasonable fit to the temporal variation in global mean temperature and the difference between Northern and Southern Hemisphere temperatures can provide a useful constraint on the net aerosol radiative forcing (as demonstrated, for example, by Harvey and Kaufmann, 2002; Stott et al., 2006c).

The inverse estimates summarized in Table 9.1 suggest that to be consistent with observed warming, the net aerosol forcing over the 20th century should be negative and likely ranges between –1.7 and –0.1 W/m^2. This assessment accounts for the probability of other forcings projecting onto the fingerprints. These results therefore typically provide a somewhat smaller upper limit for the total aerosol forcing than the estimates given in Chapter 2, which are derived from forward calculations and range between –2.2 to –0.5 W/m^2 (5% to 95%, median of –1.3 W/m^2). Note that the uncertainty ranges from inverse and forward calculations will be different due to the use of different information, and that they will be affected by different uncertainties.

Nevertheless, the similarity between results from inverse and forward estimates of aerosol forcing strengthens confidence in estimates of total aerosol forcing, despite remaining uncertainties. Harvey and Kaufmann (2002), who use an approach that focuses on the IPCC TAR range of climate sensitivity, further conclude that global mean forcing from fossil fuel-related aerosols is probably less than –1.0 W/m^2 in 1990 and that global mean forcing from biomass burning and anthropogenically-enhanced soil dust aerosols is “unlikely” to have exceeded –0.5 W/m^2 in 1990.

9.2.1.3 Radiative Forcing of Preindustrial Climate Change

Here we briefly discuss the radiative forcing estimates used for understanding climate during the last millennium, the mid-Holocene, and the Last Glacial Maximum (LGM) (Section 9.3) and in estimates of climate sensitivity based on paleoclimatic records (Section 9.6.2).

Regular variation in the Earth’s orbital parameters has been identified as the pacemaker of climate change on the glacial to interglacial timescale (see Berger, 1988 for a review). These orbital variations, which can be calculated from astronomical laws (Berger, 1978), force climate variations by changing the seasonal and latitudinal distribution of solar radiation (Chapter 6).

Insolation at the time of the LGM (21,000 years ago) was similar to today. Nonetheless, the LGM climate remained cold due to the presence of large ice-sheets in the Northern Hemisphere (Peltier, 1994, 2004) and reduced atmospheric CO_2 concentration (185 ppm according to recent ice core estimates, see Monnin et al., 2001). Most modelling studies of this period do not treat ice-sheet extent and elevation, and CO_2 concentration prognostically, but specify them as boundary conditions. The LGM radiative forcing from the reduced atmospheric concentrations of well mixed greenhouse gases is likely to have been about –2.8 W/m^2 (see Chapter 6, Figure 6.5). Ice-sheet albedo forcing is estimated to have caused a global mean forcing of about –3.2 W/m^2 (based on a range of several LGM simulations) and radiative forcing from increased atmospheric aerosols (dust primarily) is estimated to have been about -1 W/m^2. Therefore, the total annual and global mean radiative forcing during the LGM is likely to have been approximately –8 W m^{-2} relative to 1750 with large seasonal and geographical variations and with significant uncertainties (see Chapter 6, Section 6.4.1).

The major mid-Holocene forcing relative to the present was due to orbital perturbations that lead to large changes in the seasonal cycle of insolation. The Northern Hemisphere seasonal cycle was about 27 W/m^2 larger, whereas there was only a negligible change in NH annual mean solar forcing. For the Southern Hemisphere, the seasonal forcing was –6.5 W/m^2. In contrast, the global and annual mean net forcing was only 0.011 W/m^2.

Changes in the Earth's orbit have had little impact on annual mean insolation over the past millennium. Summer insolation decreased by 0.33 W/m^2 at 45°N over the millennium, winter insolation increased by 0.83 W/m^2 (Goosse et al., 2005), and the magnitude of the mean seasonal cycle of insolation in the NH decreased 0.4 W/m^2. Changes in insolation are also thought to have arisen from small variations in solar irradiance, although both timing and magnitude of past solar radiation fluctuations are highly uncertain (see Chapters 2 and 6; Lean et al., 2002; Gray et al., 2005; Foukal et al., 2006). For example, sunspots were generally missing from approximately 1675 to 1715 (the so-called Maunder Minimum) and thus solar irradiance is thought to have been reduced during this period. The estimated difference between present day solar irradiance cycle mean and the Maunder Minimum is 0.08% (–1.1 W/m^2 best estimate, range –0.5 to –2 W/m^2; see Chapter 2, Section 2.7.1.2.2), which corresponds to a radiative forcing of about 0.2 W/m^2, which is substantially lower than estimates used in the TAR (Chapter 2).

Natural external forcing also results from explosive volcanism that introduces aerosols into the stratosphere (Chapter 2, Section 2.7.2), leading to a global negative forcing during the year following the eruption. Several reconstructions are available for the last two millennia and have been used to force climate models (Chapter 6, Section 6.6.3). There is close agreement on the timing of large eruptions in the various compilations of historic volcanic activity, but large uncertainty in the magnitude of individual eruptions (Chapter 6, Figure 6.13). Different reconstructions identify similar periods when eruptions happened more frequently. The uncertainty in the overall amplitude of the reconstruction of volcanic forcing is also important for quantifying the influence of volcanism on temperature reconstructions over longer periods, but is difficult to quantify and may be a substantial fraction of the best estimate (e.g., Hegerl et al., 2006a).

9.2.2 Spatial and Temporal Patterns of the Response to Different Forcings and their Uncertainties

9.2.2.1 Spatial and Temporal Patterns of Response

The ability to distinguish between climate responses to different external forcing factors in observations depends on the extent to which those responses are distinct (see, for example, Section 9.4.1.4 and Appendix 9.A). Figure 9.1 illustrates the zonal average temperature response in the PCM model to several different forcing agents over the last 100 years (Santer et al., 2003b), while Figure 9.2 illustrates the zonal average temperature response in the CSIRO atmospheric model (when coupled to a simple mixed layer ocean model) to fossil fuel black carbon and organic matter, and to the combined effect of these forcings together with biomass burning aerosols (Penner et al., 2005). These figures indicate that the modelled vertical and zonal average signature of the temperature response should depend on the forcings. The major features shown in Figure 9.1 are robust to using different climate models. On the other hand, the response to black carbon forcing has not been widely examined and therefore the features in Figure 9.2 may be model dependent. Nevertheless, the response to black carbon forcings appears to be small.

Greenhouse gas forcing is expected to produce warming in the troposphere, cooling in the stratosphere, and, for transient simulations, somewhat more warming near the surface in the NH due to its larger land fraction (which has a shorter surface response time to the warming than do ocean regions) (Figure 9.1c). The spatial pattern of the transient surface temperature response to greenhouse gas forcing also typically exhibits a land-sea pattern of stronger warming over land, for the same reason (e.g., Cubasch et al., 2001). Sulphate aerosol forcing results in cooling throughout most of the globe, with greater cooling in the NH due to its higher aerosol loading (Figure 9.1e, see Chapter 2), thereby partially offsetting the greater NH greenhouse gas induced warming. The combined effect of tropospheric and stratospheric ozone forcing (Figure 9.1d) is expected to warm the troposphere, due to increases in tropospheric ozone, and cool the stratosphere, particularly at high latitudes where stratospheric ozone loss has been greatest. Greenhouse gas forcing is also expected to change the hydrological cycle worldwide, leading to disproportionally stronger increases in heavy precipitation (Chapter 10 and Section 9.5.4), while aerosol forcing can influence rainfall regionally (Section 9.5.4).

The simulated responses to natural forcing are distinct from those due to the anthropogenic forcings described above. Solar forcing results in a general warming of the atmosphere (Figure 9.1a) with a pattern of surface warming that is similar to that expected from greenhouse gas warming, but in contrast to the response to greenhouse warming, the simulated solar forced warming extends throughout the atmosphere (see, for example, Cubasch et al., 1997). A number of independent analyses have identified tropospheric changes that appear to be associated with the solar cycle (van Loon and Shea, 2000; Gleisner and Thejll, 2003; Haigh, 2003; White et al., 2003; Coughlin and Tung, 2004; Labitzke, 2004; Crooks and Gray, 2005), suggesting an overall wamer and moister troposphere during solar maximum. The peak-to-trough amplitude of the response to the solar cycle globally is approximately 0.1ºC near the surface. Such variations over the 11 year solar cycle make it is necessary to use several decades of data in detection and attribution studies. The solar cycle also affects atmospheric ozone concentrations with possible impacts on temperatures and winds in the stratosphere, and has been hypothesized to influence clouds through cosmic rays (Chapter 2, Section 2.7.1.3). Note that there is substantial uncertainty in the identification of climate response to solar cycle variations because the satellite period is short relative to the solar cycle length, and because the response is difficult to separate from internal climate variations and the response to volcanic eruptions (Gray et al., 2005).

Volcanic SO_2 emissions ejected into the stratosphere form sulfate aerosols and lead to a forcing that causes a surface and tropospheric cooling and a stratospheric warming that peaks several months after a volcanic eruption and lasts for several years. Volcanic forcing also likely leads to a response in the atmospheric circulation in boreal winter (discussed below) and a reduction in land precipitation (Robock and Liu, 1994; Broccoli et al., 2003; Gillett et al., 2004b). The response to volcanic forcing causes a net cooling over the 20th century because of variations in the frequency and intensity of volcanic eruptions. This results in stronger volcanic forcing towards the end of the 20th century than early in the 20th century. In the PCM, this increase results in a small warming in the lower stratosphere and near the surface at high latitudes, with cooling elsewhere (Figure 9.1b).

The net effect of all forcings combined is a pattern of NH temperature change near the surface that is dominated by the positive forcings (primarily greenhouse gases), and cooling in the stratosphere that results predominantly from greenhouse gas and stratospheric ozone forcing (Figure 9.1f). Results obtained with the CSIRO model (Figure 9.2) suggest that black carbon, organic matter and biomass aerosols would slightly enhance the Northern Hemisphere warming that is shown in Figure 9.1f. On the other hand, indirect aerosol forcing from fossil fuel aerosols may be larger than the direct effects that are represented in the CSIRO and PCM models, in which case the Northern Hemisphere warming could be somewhat diminished. Also, while land use change may cause substantial forcing regionally and seasonally, its forcing and response are expected to have only a small impact on large spatial scales (Section 9.3.3.3; Chapter 2, Figures 2.20 and 2.23; Chapter 7, Section 7.2.2)

[INSERT FIGURE 9.1 HERE]

[INSERT FIGURE 9.2 HERE]

The spatial signature of a climate model's response is seldom very similar to that of the forcing due, in part, to the strength of the feedbacks relative to the initial forcing. This comes about because climate system feedbacks vary spatially and because the atmospheric and ocean circulation cause a redistribution of energy over the globe. For example, sea ice albedo feedbacks tend to enhance the high latitude response of both a positive forcing, such as that by CO_2, and a negative forcing such as that by sulphate aerosol (e.g., Mitchell et al., 2001; Rotstayn and Penner, 2001). Cloud feedbacks can affect both the spatial signature of the response to a given forcing and the sign of the change in temperature relative to the sign of the radiative forcing (Chapter 8, Section 8.6). Heating by black carbon, for example, can decrease cloudiness (Ackerman et al., 2000). If the black carbon is near the surface, it may warm surface temperatures, while if it is at higher altitudes it may cool surface temperatures (Hansen et al., 1997; Penner et al., 2003). Feedbacks can also lead to differences in the response of different models to a given forcing agent, since the spatial response of a climate model to forcing depends on its representation of these feedbacks and processes. Additional factors that affect the spatial pattern of response include differences in thermal inertia between land and sea areas, and the lifetimes of the various forcing agents. Shorter-lived agents, such as aerosols, tend to have a more distinct spatial pattern of forcing, and can therefore also be expected to have some locally distinct response features.

The pattern of response to a radiative forcing can also be altered quite substantially if the atmospheric circulation is affected by the forcing. Modelling studies and data comparisons suggest that volcanic aerosols (e.g., Kirchner et al., 1999; Shindell et al., 1999; Yang and Schlesinger, 2001; Stenchikov et al., 2006), and greenhouse gas changes (e.g., Fyfe et al., 1999; Shindell et al., 1999; Rauthe et al., 2004), can alter the North Atlantic Oscillation or the Northern Annular mode. For example, volcanic eruptions, with the exception of high-latitude eruptions, are often followed by a positive phase of the NAM or NAO (e.g., Stenchikov et al., 2006) leading to Eurasian winter warming that may reduce the overall cooling effect of volcanic eruptions on annual averages, particularly over Eurasia (Perlwitz and Graf, 2001; Stenchikov et al., 2002; Shindell et al., 2003; Stenchikov et al., 2004; Oman et al., 2005; Rind et al., 2005b; Miller et al., 2006; Stenchikov et al., 2006). In contrast, NAM or NAO reponses to solar forcing vary between studies, some indicating a response, perhaps with dependence of the response on season or other conditions, and some finding no changes (Shindell et al., 2001b; Shindell et al., 2001a; Ruzmaikin and Feynman, 2002; Tourpali et al., 2003; Egorova et al., 2004; Palmer et al., 2004; Stendel et al., 2006, see also review in Gray et al., 2005).

In addition to the spatial pattern, the temporal evolution of the different forcings (Chapter 2, Figure 2.23) generally helps to distinguish between the responses to different forcings. For example, Santer et al. (1996b; 1996c) point out that a temporal pattern in the hemispheric temperature contrast would be expected in the second half of the 20th century with the Southern Hemisphere warming more than the Northern Hemisphere for the first two decades of this period and the Northern Hemisphere warming more than the Southern Hemisphere subsequently, as a result of changes in the relative strengths of the greenhouse gas and aerosol forcings. However, it should be noted that the integrating effect of the oceans (Hasselmann, 1976) results in climate responses that are more similar in time between different forcings than the forcings are to each other, and that there are substantial uncertainties in the evolution of the hemispheric temperature contrasts associated with sulfate aerosol forcing.

9.2.2.2 Aerosol Scattering and Cloud Feedback in Models and Observations

One line of observational evidence that reflective aerosol forcing has been changing over time comes from satellite observations of changes in top of atmosphere (TOA) outgoing shortwave flux. Increases in the outgoing shortwave flux can be caused by increases in reflecting aerosols, increases in clouds or a change in the vertical distribution of clouds and water vapor, or increases in surface albedo. Increases in aerosols and clouds can cause decreases in surface radiation fluxes and decreases in surface warming. There has been continuing interest in this possibility (Gilgen et al., 1998; Stanhill and Cohen, 2001; Liepert, 2002). Sometimes called "global dimming", this phenomena has reversed since about 1990 (Pinker et al., 2005; Wielicki et al., 2005; Wild et al., 2005) (Chapter 3, Section 3.4.3), but over the entire period from 1984 to 2001, surface solar radiation has increased by about 0.16 W m^{-2} yr^{-1} on average (Pinker et al., 2005). Figure 9.3 shows the TOA outgoing shortwave flux anomalies from the multi-model dataset at PCMDI, compared to that measured by the ERBS satellite (Wong et al., 2006) and inferred from ISCCP FD data (Zhang et al., 2004). The downward trend in outgoing solar radiation is consistent with the long term upward trend in surface radiation found by Pinker et al. (2005). The effect of Pinatubo in 1991 results in an increase in the outgoing shortwave flux (and a corresponding dimming at the surface) and its effect has been included in (but not all) models in the MMD. The ISCCP FD flux anomaly for the Pinatubo signal is almost 2 Wm^{-2} larger than that for ERBS, possibly due to the aliasing of the stratospheric aerosol signal into the ISCCP cloud properties. Overall, the trends from the ISCCP FD data (–0.18 with 95% confidence limits of ±0.11 Wm^{-2}/yr) and the ERBS data (–0.13 ± 0.08 W m^{-2}/yr) from 1984 to 1999 are not significantly different from each other at the 5% significance level, and are in even better agreement if one only considers tropical latitudes (Wong et al., 2006). These observations suggest that the effect of aerosols (and/or clouds) decreasing during this time period is larger than the general increasing cloudiness reported in Section 3.4.3. The model-predicted trends are also negative over this time period, but are smaller in most models than in the ERBS observations (which are considered more accurate than the ISCCP FD data). Wielicki et al. (2002) explained the observed downward trend by decreases in cloudiness, which is not well represented in the models on these decadal timescales (Chen et al., 2002; Wielicki et al., 2002).

[INSERT FIGURE 9.3 HERE]

9.2.2.3 Uncertainty in the Spatial Pattern of Response

Most detection methods identify the magnitude of the space-time patterns of response to forcing (sometimes called "fingerprints") that provide the best fit to the observations. The fingerprints are typically estimated from ensembles of climate model simulations forced with reconstructions of past forcing. Using different forcing reconstructions and climate models in such studies provides some indication of forcing and model uncertainty. However, few studies have examined how uncertainties in the spatial pattern of forcing explicitly contribute to uncertainties in the spatial pattern of the response. For short-lived components, uncertainties in the spatial pattern of forcing are related to uncertainties in emissions patterns, uncertainties in the transport within the climate model or chemical transport model and, especially for aerosols, uncertainties in the representation of relative humidities or clouds. These uncertainties affect the spatial pattern of the forcing. For example, the ratio of the Southern Hemisphere to Northern Hemisphere indirect aerosol forcing associated with the first aerosol indirect forcing ranges from –0.12 to 0.63 (best guess 0.29) in different studies, and that between ocean and land forcing ranges from 0.03 to 1.85 (see Chapter 7, Figure 7.5.3; Rotstayn and Penner, 2001; Chuang et al., 2002; Kristjansson, 2002; Lohmann and Lesins, 2002; Menon et al., 2002b; Rotstayn and Liu, 2003; Lohmann and Feichter, 2005).

9.2.2.4 Uncertainty in the Temporal Pattern of Response

Climate model studies have also not systematically explored the effect of uncertainties in the time-evolution of forcings. These uncertainties depend mainly on the uncertainty in the spatio-temporal expression of emissions, and, for some forcings, fundamental understanding of the possible change over time.

The increasing forcing by greenhouse gases is relatively well known. Also, the global time history of SO_2 emissions, which has a larger overall forcing than that of the other short-lived aerosol components, is quite well constrained. Seven different reconstructions of the time history of global anthropogenic sulfur emissions up to 1990 have a relative standard deviation of less than 20% between 1890 and 1990, with better agreement in more recent years. This robust time history increases confidence in results from detection and attribution studies that attempt to separate the effects of sulfate aerosol and greenhouse gas forcing (Section 9.4.1).

In contrast, there are large uncertainties related to the anthropogenic emissions of other short-lived compounds and their effects on forcing. For example, estimates of historical emissions from fossil fuel combustion do not account for changes in emission factors (the ratio of the emitted gas or aerosol to the fuel burned) of short-lived species associated with concerns over urban air pollution (e.g., van Aardenne et al., 2001). Changes in these emission factors would have slowed the emissions of NOx as well as CO after about 1970 and slowed the accompanying increase of tropospheric ozone compared to that represented by a single emission factor for fossil fuel use. In addition, changes in the height of emissions of SO_2 associated with the implementation of tall stacks would have changed the lifetime of sulfate aerosols and the relationship between emissions and effects. Another example relates to the emissions of black carbon associated with the burning of fossil fuels. The spatial and temporal emissions of black carbon by continent reconstructed by Ito and Penner (2005) are significantly different from those reconstructed using the methodology of Novakov et al. (2003). For example, the emissions in Asia grow significantly faster in the inventory based on Novakov et al. (2003) compared to those based on Ito and Penner (2005). Also, before 1988 the growth in emissions in Eastern Europe using the Ito and Penner (2005) inventory is faster than the growth based on the methodology of Novakov et al. (2003). Such spatial/temporal uncertainties will contribute to both spatial/temporal uncertainties in the net forcing and to spatial/temporal uncertainties in the distribution of forcing and response.

There are also large uncertainties in the magnitude of low-frequency changes in forcing associated with total solar radiation changes as well as its spectral variation, particularly on timescales longer than the 11-year cycle. Previous estimates of change in the total solar radiation have used sun spot numbers to calculate these slow changes in solar irradiance over the last few centuries, but these earlier estimates are not necessarily supported by current understanding and the estimated magnitude of low-frequency changes has been substantially reduced since the TAR (Lean et al., 2002; Foukal et al., 2004; Foukal et al., 2006; Chapter 6, Section 6.4.1; Chapter 2, Section 2.7.1.2). In addition, the magnitude of radiative forcing associated with major volcanic eruptions is uncertain and differs between reconstructions (Sato et al., 1993; Andronova et al., 1999; Ammann et al., 2003), although the timing of the eruptions is well documented.

9.2.3 Implications for Understanding 20th Century Climate Change

Any assessment of observed climate change that compares simulated and observed responses will be affected by errors and uncertainties in the forcings prescribed in a climate model and its corresponding responses. As noted above, detection studies scale the response patterns to different forcings to obtain the best match to observations. Thus errors in the magnitude of the forcing or in the magnitude of the model response to a forcing (which is approximately, although not exactly, a function of climate sensitivity), should not affect detection results provided that the large-scale space-time pattern of the response is correct. Attribution studies evaluate the consistency between the model-simulated amplitude of response and that which is inferred from observations. In the case of uncertain forcings, scaling factors provide information about strength of the forcing (and response) needed to reproduce the observations, or about the possibility that the simulated pattern or strength of response is incorrect. However, for a model simulation to be considered consistent with the observations given forcing uncertainty, the forcing used in the model should remain consistent with the uncertainty bounds from forward model estimates of forcing.

Detection and attribution approaches that try to distinguish the response to several external forcings simultaneously may be affected by similarities in the pattern of response to different forcings and by uncertainties in forcing and response. Similarities between the responses to different forcings, particularly in the spatial patterns of response, make it more difficult to distinguish between responses to different external forcings, but also imply that the response patterns will be relatively insensitive to modest errors in the magnitude and distribution of the forcing. Differences between the time histories of different kinds of forcing (e.g., greenhouse gas versus sulphate aerosol) ameliorate the problem of the similarity between the spatial patterns of response considerably. For example, the spatial response of surface temperature to solar forcing resembles that due to anthropogenic greenhouse gases forcing (Weatherall and Manabe, 1975; Nesme-Ribes et al., 1993; Cubasch et al., 1997; Rind et al., 2004; Zorita et al., 2005). Distinct features of the vertical structure of the responses in the atmosphere to different types of forcing further help to distinguish between the different sources of forcing. Studies that interpret observed climate in subsequent sections use such strategies, and the overall assessment in this chapter uses results from a range of climate variables and observations.

Many detection studies attempt to identify in observations both temporal and spatial aspects of the temperature response to a given set of forcings because the combined space-time responses tend to be more distinct than either the space-only or time-only patterns of response. Because the emissions and burdens of different forcing agents change with time, the net forcing and its rate of change vary with time. Although explicit accounting for uncertainties in the net forcing is not available (see discussion in Sections 9.2.2.3, 9.2.2.4), models often employ different implementations of external forcing. Detection and attribution studies using such simulations suggest that results are not very sensitive to moderate forcing uncertainties. A further problem arises due to spurious temporal correlations between the responses to different forcings that arise from sampling variability. For example, spurious correlation between the climate responses to solar and volcanic forcing over parts of the 20th century (North and Stevens, 1998) can lead to misidentification of one as the other, as in Douglass and Clader (2002).

The spatial pattern of the temperature response to aerosol forcing is quite distinct from the spatial response pattern to CO_2 in some models and diagnostics (Hegerl et al., 1997), but less so in others (Reader and Boer, 1998; Tett et al., 1999; Hegerl et al., 2000; Harvey, 2004). If it is not possible to distinguish the spatial pattern of greenhouse warming from that of fossil-fuel related aerosol cooling, the observed warming over the last century could be explained by large greenhouse warming balanced by large aerosol cooling or alternatively by small greenhouse warming with very little or no aerosol cooling. Nevertheless, estimates of the amplitude of the response to greenhouse forcing in the 20^{th} century from detection studies are quite similar, even though the simulated responses to aerosol forcing are model dependent (Gillett et al., 2002a; Hegerl and Allen, 2002). By considering three different climate models, Stott et al. (2006c) concluded that an important constraint on the possible range of responses to aerosol forcing is the temporal evolution of the global mean and hemispheric temperature contrast as was suggested by Santer et al. (1996a). See also Sections 9.2.4 and 9.4.1.5.

9.2.4 Summary

The uncertainty in the magnitude and spatial pattern of forcing differs considerably between forcings. For example, well-mixed greenhouse gas forcing is relatively well constrained and spatially homogeneous. In contrast, uncertainties are large for many non-greenhouse gas forcings. Inverse model studies, which use methods closely related to those used in climate change detection research, indicate that the magnitude of the total net aerosol forcing has a likely range of –1.7 to –0.1 to W/m^2. As summarized in Chapter 2, forward calculations of aerosol radiative forcing, which do not depend on knowledge of observed climate change or the ability of climate models to simulate the transient response to forcings, provide results (–2.2 to –0.5 W m^{-2}; 5 to 95%) that are quite consistent with inverse estimates; the uncertainty ranges from inverse and forward calculations will be different due to the use of different information. The large uncertainty in total aerosol forcing makes it more difficult to accurately infer the climate sensitivity from observations (Section 9.6). It also increases uncertainties in results that attribute cause to observed climate change (Section 9.4.1.4), and is in part responsible for differences in probabilistic projections of future climate change (Chapter 10). Forcings from black carbon, fossil fuel organic matter, and biomass burning aerosols, which have not been considered in most detection studies performed to date, are likely small but with large uncertainties relative to the magnitudes of the forcings.

Uncertainties also differ between natural forcings and sometimes between different time scales for the same forcing. For example, while the 11-year solar forcing cycle is well-documented, lower-frequency variations in solar forcing are highly uncertain. Furthermore, the physics of the response to solar forcing and some feedbacks are still poorly understood. In contrast, the timing and duration of forcing due to aerosols ejected into the stratosphere by large volcanic eruptions is well known during the instrumental period, although the magnitude of that forcing is uncertain.

Differences in the time-evolution and sometimes the spatial pattern of climate response to external forcing make it possible, with limitations, to separate the response to these forcings in observations, such as the responses to greenhouse gas and sulphate aerosol forcing. In contrast, the climate response and time evolution of other anthropogenic forcings is more uncertain, making the simulation of the climate response and its detection in observations more difficult. The time-evolution, and to some extent the spatial and vertical pattern, of the climate response to natural forcings is also quite different from that of anthropogenic forcing. This makes it possible to separate the climate response to solar and volcanic forcing from that to anthropogenic forcing despite the uncertainty in the history of solar forcing noted above.

9.3 Understanding Pre-Industrial Climate Change

9.3.1 Why Consider Pre-Industrial Climate Change?

The Earth system has experienced large-scale climate changes in the past (Chapter 6) that hold important lessons for the understanding of present and future climate change. These changes resulted from natural external forcings that, in some instances, triggered strong feedbacks as in the case of the Last Glacial Maximum (LGM, see Chapter 6). Past periods offer the potential to provide information not available from the instrumental record, which is affected by anthropogenic as well as natural external forcings and is too short to fully understand climate variability and major climate system feedbacks on interdecadal and longer timescales. Indirect indicators ("proxy data" such as tree ring width and density) must be used to infer climate variations (Chapter 6) prior to the instrumental era (Chapter 3). A complete description of these data and of their uncertainties can be found in Chapter 6.

The discussion here is restricted to several periods in the past for which modelling and observational evidence can be compared to test our understanding of the climate response to external forcings. One such period is the last millennium, which places the recent instrumental record in a broader context (e.g., Mitchell et al., 2001). The analysis of the past 1000 years focuses mainly on the climate response to natural forcings (changes in solar radiation and volcanism) and on the role of the anthropogenic forcing during the most recent part of the record. We also consider two time periods analyzed in the Paleoclimate Model Intercomparison Project (PMIP, Joussaume and Taylor, 1995; PMIP2, Harrison et al., 2002), the mid-Holocene (6000 years ago) and last Glacial Maximum (21000 years ago). Both periods had a substantially different climate compared to the present, and there is relatively good information from data synthesis and model simulation experiments (Braconnot et al., 2004; Cane et al., 2006). An increased number of simulations using models of intermediate complexity or AOGCMs that are the same as, or related to, those used in simulations of the climates of the 20th and 21st centuries are available for these periods.

9.3.2 What can be learned from the Last Glacial Maximum and the Mid-Holocene?

Relatively high quality global terrestrial climate reconstructions exist for last glacial maximum (LGM) and the mid-Holocene and as part of the BIOME6000 project (Prentice and Webb, 1998; Prentice and Jolly, 2000). The CLIMAP (1981) reconstruction of LGM sea surface temperatures has also been improved (Chapter 6). The LGM climate was colder and drier than at present as is indicated by the extensive tundra and steppe vegetation that existed during this period. Most LGM proxy data suggest that the tropical oceans were colder by about 2°C than at present, and that the frontal zones in the Southern and Northern Hemispheres were shifted equator wards (Kucera et al., 2005), even though large differences are found between temperature estimates from the different proxies in the North Atlantic.

Several new AOGCM simulations of the LGM have been produced since the TAR. These simulations show a global cooling of approximately 3.5–5.2°C when LGM greenhouse gas and ice sheet boundary conditions are specified (Chapter 6), which is within the range (–1.8°C to –6.5°C) of PMIP results from simpler models that were discussed in the TAR (McAvaney et al., 2001). Only one simulation exhibits a very strong response with a cooling of approximately 10°C (Kim et al., 2002). All of these simulations exhibit a strongly damped hydrological cycle relative to that of the modern climate, with less evaporation over the oceans and continental scale drying over land. Changes in greenhouse gas concentrations may account for about half of the simulated tropical cooling (Shin et al., 2003), and for the production of colder and saltier water found at depth in the Southern Ocean (Liu et al., 2005). Most LGM simulations with coupled models shift the deep-water formation in the North Atlantic southward, but large differences exist between models in the intensity of the Atlantic meridional overturning. Including vegetation changes appears to improve the realism of LGM simulations (Wyputta and McAvaney, 2001). Furthermore, including the physiological effect of the CO_2 concentration on vegetation has a non-negligible impact (Levis et al., 1999) and is necessary to properly represent changes in global forest (Harrison and Prentice, 2003) and terrestrial carbon storage (e.g., Kaplan et al., 2002; Joos et al., 2004, see also Chapter 6). To summarize, despite large uncertainties, LGM simulations capture the broad features found in paleo-climate data, and better agreement is obtained with new coupled simulations using more recent models and more complete feedbacks from ocean, sea-ice and land surface characteristics such as vegetation and soil moisture (Chapter 6).

Closer to the present, during the mid-Holocene, one of the most noticeable indications of climate change is the northward extension of northern temperate forest (Bigelow et al., 2003), which reflects warmer summers than at present. In the tropics the more vegetated conditions inferred from pollen records in the now dry sub-Saharan regions indicate wetter conditions due to enhanced summer monsoons (see Braconnot et al., 2004 for a review). AOGCM simulations of the mid-Holocene (see Section 9.2.1.3 for forcing) produce an amplification of the mean seasonal cycle of temperature of approximately 0.5°C to 0.7°C. This range is slightly smaller than that obtained using atmosphere only models in PMIP1 (~0.5°C to ~1.2°C) due to the thermal response of the ocean (Braconnot et al., 2000). Simulated changes in the ocean circulation have strong seasonal features with an amplification of the SST seasonal cycle of 1–2°C in most places within the tropics (Zhao et al., 2005), influencing the Indian and African monsoons. Over West Africa, AOGCM simulated changes in annual mean precipitation are about 5–10% larger than for atmosphere only simulations, and in better agreement with data reconstructions (Braconnot et al., 2004). Results for the Indian and Southwest Asian monsoon are less consistent between models.

As noted in the TAR (McAvaney et al., 2001), vegetation change during the mid-Holocene likely triggered changes in the hydrological cycle, explaining the wet conditions that prevailed in the Sahel region, that were further enhanced by ocean feedbacks (Ganopolski et al., 1998; Braconnot et al., 1999), although soil moisture may have counteracted some of these feedbacks (Levis et al., 2004). Wohlfahrt et al. (2004) showed that in mid- and high-latitudes the vegetation and ocean feedbacks enhanced the warming in spring and autumn by about 0.8°C. However, models have a tendency to overestimate the mid-continental drying in Eurasia, which is further amplified when vegetation feedbacks are included (Wohlfahrt et al., 2004).

A wide range of proxies containing information about ENSO variability during the mid-Holocene is now also available (Chapter 6, Section 6.5.3). These data suggest that ENSO variability was weaker than today prior to approximately 5,000 years before present (Moy et al., 2002 and references therein; Tudhope and Collins, 2003). Several studies have attempted to analyse these changes in interannual variability from model simulations. Even though some results are controversial, a consistent picture has emerged for the mid-Holocene, for which simulations produce reduced variability in precipitation over most ocean regions in the tropics (Liu et al., 2000; Braconnot et al., 2004; Zhao et al., 2005). Results obtained with the Cane-Zebiak model suggest that the Bjerknes (1969) feedback mechanism may be a key element of the ENSO response in that model. The increased mid-Holocene solar heating in boreal summer leads to more warming in the western than eastern Pacific, which strengthens the trade winds and inhibits the development of ENSO (Clement et al., 2000; Clement et al., 2004). AOGCMs also tend to simulate less intense ENSO events, in qualitative agreement with data, although there are large differences in magnitude and proposed mechanisms and inconsistent responses of the associated teleconnections (Otto-Bliesner, 1999; Liu et al., 2000; Kitoh and Murakami, 2002; Otto-Bliesner et al., 2003).

9.3.3 What can be learned from the Past 1000 Years?

External forcing relative to the present is generally small for the last millennium when compared to that for the mid-Holocene and LGM. Nonetheless, there is evidence that climatic responses to forcing, together with natural internal variability of the climate system, produced several well-defined climatic events, such as the cool conditions during the 17th century or relatively warm periods early in the millennium.

9.3.3.1 Evidence of External Influence on the Climate Over the Past 1000 years

A substantial number of proxy reconstructions of annual or decadal Northern Hemisphere mean surface temperature are now available (see Chapter 6, Figure 6.11, and the reviews by Jones et al., 2001 and Jones and Mann, 2004). Several new reconstructions have been published, some of which suggest larger variations over the last millennium than assessed in the TAR, but uncertainty remains in the magnitude of inter-decadal to inter-century scale variability. This uncertainty arises because different studies rely on different proxy data or use different reconstruction methods (Chapter 6, Section 6.6.1). Nonetheless, Northern Hemisphere mean temperatures in the second half of the 20th century were likely warmer than in any other 50-year period in the last 1300 years (Chapter 6), and very likely warmer than any such period in the last 500 years. Temperatures decreased subsequently, and then rose rapidly during the most recent hundred years. This long term tendency is punctuated by substantial shorter term variability (Chapter 6, Figure 6.10). For example, cooler conditions with temperatures 0.5–1°C below the 20th century mean value are found in the 17th and early 18th centuries.

A number of simulations of the last millennium (Chapter 6, Figure 6.13) have been performed using a range of models, including some simulations with AOGCMs (e.g., Crowley, 2000; Goosse and Renssen, 2001; Bertrand et al., 2002; Bauer et al., 2003; Gerber et al., 2003; see also Gonzalez-Rouco et al., 2003; Jones and Mann, 2004; Zorita et al., 2004; Weber, 2005; Tett et al., 2006). These simulations use different reconstructions of external forcing, particularly solar, volcanic and greenhouse gas forcing, and often also include land use changes (e.g., Bertrand et al., 2002; Stendel et al., 2006; Tett et al., 2006). While the use of different models and forcing reconstructions leads to differences, the simulated evolution of the Northern Hemisphere annual mean surface temperature displays some common characteristics between models that are consistent with the broad features of the data (Chapter 6, Figure 6.13 and Figure 9.4). For example, all simulations show relatively cold conditions during the period around 1675 to 1715 in response to natural forcing, which is in qualitative agreement with the proxy reconstructions. In all simulations shown in Chapter 6, Figure 6.13 the late 20th century is warmer than any other multi-decadal period during the last millennium. Also, there is significant correlation between simulated and reconstructed variability (e.g., Yoshimori et al., 2005) By comparing simulated and observed atmospheric CO_2 concentration during the last 1000 years, Gerber et al. (2003) suggest that the amplitude of the temperature evolution simulated by simple climate models and EMICs is consistent with the observed evolution of CO_2. Since reconstructions of external forcing are virtually independent from the reconstructions of past temperatures, this broad consistency increases confidence in the broad features of the reconstructions and the understanding of the role of external forcing in recent climate variability. The simulations also show that it is not possible to reproduce the large 20th century warming without anthropogenic forcing regardless of which solar or volcanic forcing reconstruction is used (Crowley, 2000; Bertrand et al., 2002; Bauer et al., 2003; Hegerl et al., 2003; Hegerl et al., 2006b) stressing the impact of human activity on the recent warming.

While there is broad qualitative agreement between simulated and reconstructed temperatures, it is difficult to fully assess model simulated variability because of uncertainty in the magnitude of historical variations in the reconstructions and differences in the sensitivity to external forcing (Chapter 8, Table 8.2). The role of internal variability has been found to be smaller than that of the forced variability for hemispheric temperature means at decadal or longer time scales (Crowley, 2000; Hegerl et al., 2003; Goosse et al., 2004; Weber et al., 2004; Hegerl et al., 2006b; Tett et al., 2006), and thus internal variability is a relative small contributor to differences between different simulations of Northern Hemisphere mean temperature. Other sources of uncertainty in simulations include model ocean initial conditions which, for example, explain the warm conditions found in the Zorita et al. (2004) simulation during the first part of the millennium (Goosse et al., 2005; Osborn et al., 2006).

9.3.3.2 Role of Volcanism and Solar Irradiance

Volcanic eruptions cause rapid decreases in hemispheric and global mean temperatures followed by gradual recovery over several years (Section 9.2.2.1) in climate simulations driven by volcanic forcing (Chapter 6, Figure 6.13; Crowley, 2000; Bertrand et al., 2002; Weber, 2005; Yoshimori et al., 2005; Tett et al., 2006). These simulated changes appear to correspond to cool episodes in proxy reconstructions (Chapter 6, Figure 6.13). This suggestive correspondence has been confirmed in comparisons between composites of temperatures following multiple volcanic eruptions in simulations and reconstructions (Hegerl et al., 2003;

Weber, 2005). In addition, changes in the frequency of large eruptions result in climate variability on decadal and possibly longer time-scales (Crowley, 2000; Briffa et al., 2001; Bertrand et al., 2002; Bauer et al., 2003; Weber, 2005). Hegerl et al. (2003; 2006b), using a multi-regression approach based on EBM simulated fingerprints of solar, volcanic and greenhouse gas forcing (Appendix 9.A.1; see also Section 9.4.1.4 for the 20th century), simultaneously detect the responses to volcanic and greenhouse gas forcing in a number of proxy reconstructions of average Northern Hemisphere mean annual and growing season temperatures (Figure 9.4) with high significance. They find that a high percentage of decadal variance in the reconstructions used can be explained by external forcing (between 49% and 70% of decadal variance depending upon the reconstruction).

[INSERT FIGURE 9.4 HERE]

There is more uncertainty regarding the influence of solar forcing. In addition to substantial uncertainty in the timing and amplitude of solar variations on timescales of several decades to centuries, which has increased since the TAR, although the estimate of solar forcing has been revised downwards (Section 9.2.1.3 and Chapter 2, Section 2.7.1), uncertainty also arises because the spatial response of surface temperature to solar forcing resembles that due to greenhouse gas forcing (Section 9.2.3). Analyses that make use of differences in the temporal evolution of solar and volcanic forcings are better able to distinguish between the two (Section 9.2.3; see also Section 9.4.1.5 for the 20th century). In such an analysis, solar forcing can only be detected and distinguished from the effect of volcanic and greenhouse gas forcing over some periods in some reconstructions (Hegerl et al., 2003; 2006b), although the effect of solar forcing has been detected over parts of the 20th century in some time-space analyses (Section 9.4.1.5) and there are similarities between regressions of solar forcing on model simulations and several proxy reconstructions (Weber, 2005, see also Waple, 2002). A model simulation (Shindell et al., 2003) suggests that solar forcing may play a substantial role for regional anomalies due to dynamical feedbacks. These uncertainties in the contribution of different forcings to climatic events during the last millennium reflect substantial uncertainty in knowledge about past solar and volcanic forcing, as well as differences in the way these effects are taken into account in model simulations.

Overall, modelling and detection and attribution studies confirm a role of volcanic, greenhouse gas and probably solar forcing in explaining the broad temperature evolution of the last millennium, although the role of solar forcing has recently been questioned (Foukal et al., 2006). The variability that remains in proxy reconstructions after estimates of the responses to external forcing have been removed is broadly consistent with AOGCM simulated internal variability (e.g., Hegerl et al., 2003; Hegerl et al., 2006b), providing a useful check on AOGCMs even though uncertainties are large. Such studies also help to explain episodes of the climate of the last millennium. For example, several modelling studies suggest that volcanic activity has a dominant role in explaining the cold conditions that prevailed from 1675 to 1715 (Andronova et al., 2005; Yoshimori et al., 2005). In contrast, Rind et al. (2004) estimated from model simulations that the cooling relative to today was primarily associated with reduced greenhouse gas forcing, with a substantial contribution from solar forcing.

There is also some evidence from proxy data that the response to external forcing may influence modes of climate variability. For example, Cobb et al. (2003), using fossil corals, attempted to extend the ENSO record back through the last millennium. They find that ENSO events may have been as frequent and intense during the mid-seventeenth century as during the instrumental period, with events possibly rivalling the strong 1997–1998 event. On the other hand, there are periods during the 12th and 14th centuries when there may have been significantly less ENSO variability, a period during which there were also cooler conditions in the North East Pacific (MacDonald and Case, 2005) and evidence of droughts in central North America (Cook et al., 2004). Cobb et al. (2003) found that fluctuations in reconstructed ENSO variability do not appear to be correlated in an obvious way with mean state changes in the tropical Pacific or global mean climate, while Adams et al. (2003) found statistical evidence for an El Niño-like anomaly during the first few years following explosive tropical volcanic eruptions. The Cane-Zebiak model simulates changes similar to those n the Cobb et al. (2003) data when volcanism and solar forcing are accounted for, supporting the link with volcanic forcing over the past millennium (Mann et al., 2005). However, additional studies with different models are needed to fully assess this relationship, since previous work was less conclusive (Robock, 2000).

Extratropical variability also appears to respond to volcanic forcing. During the winter following a large volcanic eruption, the zonal circulation may be more intense, causing a relative warming over the continents during the cold season that could partly offset the direct cooling due to the volcanic aerosols (Section 9.2.1.1 and 8.4.1; Robock, 2000; Shindell et al., 2003). A tendency towards the negative NAO state during periods of reduced solar input is found in some reconstructions of this pattern for the northern hemisphere (Shindell et al., 2001b; Luterbacher et al., 2002; Luterbacher et al., 2004; Stendel et al., 2006), possibly implying a solar forcing role in some long-term regional changes, such as the cooling over the Northern Hemispheric continents around 1700 (Shindell et al., 2001b; Section 9.2.2). Indications of changes in ENSO variability during the low solar irradiance period of the 17th to early 18th centuries are more controversial (e.g., D'Arrigo et al., 2005).

9.3.3.3 Other Forcings and Sources of Uncertainties

In addition to forcing uncertainties discussed above, there are a number of other uncertainties that affect the understanding of preindustrial climate change. For example, land cover change may have influenced the preindustrial climate (Bertrand et al., 2002; Bauer et al., 2003), leading to a regional cooling of 1–2°C in winter and spring over the major agricultural regions of North America and Eurasia in some model simulations, when pre-agriculture vegetation was replaced by present-day vegetation (Betts, 2001). The largest anthropogenic land cover changes involve deforestation (Chapter 2). The greatest proportion of deforestation has occurred in the temperate regions of the Northern Hemisphere (Ramankutty and Foley, 1999; Goldewijk, 2001), when Europe had cleared about 80% of its agricultural area by 1860, but over half of the forest removal in North America took place since 1860 (Betts, 2001), mainly in the late 19th century (Stendel et al., 2006). During the past two decades the CO_2 flux caused by land use changes has been dominated by tropical deforestation (Chapter 7, Section 7.3.2.1.2). Climate model simulations suggest that the effect of land use change was likely small on hemispheric and global scales, estimated variously as: –0.02K relative to natural pre-agricultural vegetation (Betts, 2001), less than –0.1K since 1700 (Stendel et al., 2006), and about –0.05K over the twentieth century and too small to be detected statistically in observed trends (Matthews et al., 2004). However, the same authors did find a larger cooling effect since 1700 of between –0.06 and –0.22K when they explored the sensitivity to different representations of land cover change.

Oceanic processes and ocean-atmospheric interaction may also have played a role in the climate evolution during the last millennium (Delworth and Knutson, 2000; Weber et al., 2004; van der Schrier and Barkmeijer, 2005). Climate models generally simulate a weak to moderate increase in the intensity of the oceanic meridional overturning circulation in response to a decrease in solar irradiance (Cubasch et al., 1997; Goosse and Renssen, 2004; Weber et al., 2004). A delayed response to natural forcing due to the storage and transport of heat anomalies by the deep ocean has been proposed to explain the warm Sourthern Ocean around the 14–15th centuries (Goosse et al., 2004).

9.3.4 Summary

Considerable progress has been made since the TAR in understanding the response of the climate system to external forcings. Periods like the mid-Holocene and the Last Glacial Maximum are now used as benchmarks for climate models that are used to simulate future climate (Chapter 6). While considerable uncertainties remain in the climate reconstructions for these periods, and in the boundary conditions used to force climate models, comparisons between simulated and reconstructed conditions in the LGM and Mid Holocene demonstrate that models capture the broad features of changes in the temperature and precipitation patterns. These studies have also increased understanding of the roles of ocean and vegetation feedbacks in determining the response to solar and greenhouse gases forcing. Moreover, although proxy data on paleo-climate interannual to multidecadal variability during these periods remain very uncertain, there is an increased appreciation that external forcing may, in the past, have affected climatic variability such as that associated with ENSO.

The understanding of climate variability and change, and its causes during the past thousand years has also improved since the TAR (IPCC, 2001). There is consensus across all millennial reconstructions on the timing of major climatic events, although their magnitude remains somewhat uncertain. Nonetheless, the larger and more closely scrutinized collection of reconstructions from paleo data than were available for the TAR indicate that it is likely that Northern Hemispheric average temperatures during the second half of the

20th century twere warmer than any other 50-year period during the past 1300 years (Chapter 6). While uncertainties remain in temperature and forcing reconstructions, and in the models used to estimate the responses to external forcings, the available detection studies, modelling and other evidence support the conclusion that volcanic and possibly solar forcings have very likely affected Northern Hemisphere mean temperature over the past millennium and that external influences explain a substantial fraction of interdecadal temperature variability in the past. The available evidence also indicates that natural forcing may have influenced the climatic conditions of individual periods, such as the cooler conditions around 1700. The climate response to greenhouse gas increases can be detected in a range of proxy reconstructions by the end of the records.

AOGCMs, when driven with estimates of external forcing for the last millennium, simulate changes in hemispheric mean temperature that are in broad agreement with proxy reconstructions (given their uncertainties), increasing confidence in the forcing reconstructions, proxy climate reconstructions and models. In addition, the residual variability in the proxy climate reconstructions that is not explained by forcing is broadly consistent with AOGCM simulated internal variability. Overall, the information on temperature change over the last millennium is broadly consistent with the understanding of climate change in the instrumental era.

9.4 Understanding of Air Temperature Change During the Industrial Era

9.4.1 Global Scale Surface Temperature Change

9.4.1.1 Observed Changes

There have been six more years of observations since the TAR (Chapter 3) that show that temperatures are continuing to warm near the surface of the planet. The annual global mean temperature for every year since the TAR has been amongst the 10 warmest years since the beginning of the instrumental record. The global mean temperature averaged over land and ocean surfaces warmed by 0.76 ± 0.19°C between the first 50 years of the instrumental record (1850–1899) and the last 5 years (2001–2005) (Chapter 3; with a linear warming trend of 0.74 ± 0.18°C over the last 100 years (1906–2005)). The rate of warming over the last 50 years is almost double that over the last 100 years (0.13 ± 0.03 vs 0.07 ± 0.02°C decade^{-1}; Chapter 3). The larger number of proxy reconstructions from paleo data than were available for the TAR indicate that it is very likely that average Northern Hemisphere temperatures during the second half of the 20th century were warmer than any other 50-year period in the last 500 years and it is likely that this was the warmest period in the past 1000 years and unusually warm compared with the last 1300 years (Chapter 6). Global mean temperature has not increased smoothly since 1900 as would be expected if it were influenced only by forcing from increasing greenhouse gas concentrations (i.e., if natural variability and other forcings did not have a role; see Section 9.2,1, Chapter 2). A rise in near-surface temperatures also occurred over several decades during the first half of the 20th century, and in between there was a period of more than three decades when temperatures showed no pronounced trend (Chapter 3, Figure 3.6). Since the mid-1970s, land regions have warmed at a faster rate than oceans in both hemispheres (Chapter 3, Figure 3.8) and warming over the Southern Hemisphere was smaller than that over the Northern Hemisphere during this period (Chapter 3, Figure 3.6), while warming rates during the early 20th century were similar over land and ocean.

9.4.1.2 Simulations of the 20th Century

There are now a greater number of climate simulations from AOGCMs for the period of the global surface instrumental record than were available for the TAR, including a greater variety of forcings in a greater variety of combinations. These simulations used models with different climate sensitivities, rates of ocean heat uptake and magnitudes and types of forcings (Supplementary Material, Table S9.1). Figure 9.5 shows that simulations that incorporate anthropogenic forcings, including increasing greenhouse gas concentrations and the effects of aerosols, and that also incorporate natural external forcings provide a consistent explanation of the observed temperature record, whereas simulations that include only natural forcings do not simulate the warming observed over the last three decades. A variety of different forcings are used in these simulations. For example, some anthropogenically forced simulations include both the direct and indirect effects of sulphate aerosols whereas others include just the direct effect, and the aerosol forcing that is calculated within models differs due to differences in the representation of physics. Similarly, the effects of tropospheric and stratospheric ozone changes are included in some simulations but not others, and a few simulations include the effects of carbonaceous aerosols and land use changes, whilst the naturally forced simulations include a variety of different representations of changing solar and volcanic forcing. Despite this additional uncertainty there is a clear separation in Figure 9.5 between the simulations with anthropogenic forcings and those without.

Global mean and hemispheric scale temperatures on multi-decadal time scales are largely controlled by external forcings (Stott et al., 2000). This external control is demonstrated by ensembles of model simulations with identical forcings (whether anthropogenic or natural) whose members exhibit very similar simulations of global mean temperature on multi-decadal timescales (e.g., Stott et al., 2000; Broccoli et al., 2003; Meehl et al., 2004). Larger inter-annual variations are seen in the observations than in the ensemble mean model simulation of the 20th century because the ensemble averaging process filters out much of the natural internal inter-annual variability that is simulated by the models. The interannual variability in the individual simulations that is evident in Figure 9.5 suggests that current models generally simulate large scale natural internal variability quite well, and also capture the cooling associated with volcanic eruptions on shorter timescales. Section 9.4.1.3 will assess the variability of near surface temperature observations and simulations.

[INSERT FIGURE 9.5 HERE]

The fact that climate models are only able to reproduce observed global mean temperature changes over the 20th century when they include anthropogenic forcings, and that they fail to do so when they exclude anthropogenic forcings, is evidence for the influence of humans on global climate. Further evidence is provided by spatial patterns of temperature change. Figure 9.6 compares observed near-surface temperature trends over the globe (top row) with those simulated by climate models when they include anthropogenic and natural forcing (2nd row) and the same trends simulated by climate models when only natural forcings are included (3rd row). The observed trend over the entire 20th century (Figure 9.6 top left panel) shows warming almost everywhere with the exception of the southeastern United States, northern North Atlantic, and isolated gridboxes in Africa and South America; see also Figure 3.9. Such a pattern of warming is not associated with known modes of internal climate variability. For example, while El-Niño or El-Niño like decadal variability results in unusually warm annual temperatures, the spatial pattern associated with such a warming is more structured with cooling in the North Pacific and South Pacific (see, for example, Zhang et al., 1997). In contrast, the trends in climate model simulations that include anthropogenic and natural forcing (Figure 9.6, 2nd row) show a similar pattern of spatially near-uniform warming as observed. There is much greater similarity between the general evolution of the warming in observations and that simulated by models when anthropogenic and natural forcing is included than when only natural forcing is included (Figure 9.6, 3rd row). Figure 9.6 (4th row) shows that climate models are only able to reproduce the observed patterns of zonal-mean near-surface temperature trends over the 1901–2005 and 1979–2005 periods when they include anthropogenic forcings and fail to do so when they exclude anthropogenic forcings. Although there is less warming at low latitudes than at high northern latitudes there is also less internal variability at low latitudes which results in a greater separation of the climate simulations with and without anthropogenic forcings.

[INSERT FIGURE 9.6 HERE]

Climate simulations are consistent in showing that the global mean warming observed since 1970 can only be reproduced when models are forced with combinations of external forcings that include anthropogenic forcings (Figure 9.5). This conclusion holds despite a variety of different anthropogenic forcings and processes being included in these models (e.g., Tett et al., 2002; Broccoli et al., 2003; Meehl et al., 2004; Knutson et al., 2006). In all cases the response to forcing from well-mixed greenhouse gases dominates the anthropogenic warming in the model. No climate model using natural forcings alone has reproduced the observed global warming trend in the second half of the 20th century. Therefore, modelling studies suggest that late 20th century warming is much more likely to be anthropogenic than natural in origin, a finding which is confirmed in studies relying on formal detection and attribution methods (Section 9.4.1.4).

Modelling studies are also in moderately good agreement with observations during the first half of the 20th century when both anthropogenic and natural forcings are considered, although assessments of which forcings are important differ, with some studies finding that solar forcing is more important (Meehl et al., 2004) whilst other studies find that volcanic forcing (Broccoli et al., 2003) or internal variability (Delworth and Knutson, 2000) could be more important. Differences between simulations including greenhouse gas forcing only and those that also include the cooling effects of sulphate aerosols (e.g., Tett et al., 2002) indicate that the cooling effects of sulphate aerosols may account for some of the lack of observational warming between 1950 and 1970, despite increasing greenhouse gas concentrations, as was proposed by Schwartz et al. (1993). In contrast, Nagashima et al. (2006) find that carbonaceous aerosols are required for the MIROC model to provide a statistically consistent representation of observed changes in near-surface temperature in the middle part of the 20th century. The mid century cooling that the model simulates in some regions is also observed, and is caused in the model by regional negative surface forcing from organic and black carbon associated with biomass burning. Variations in the Atlantic Multidecadal Oscillation (see Section 3.6.6.1 for a more detailed discussion) could account for some of the evolution of global and hemispheric mean temperatures during the instrumental period (Schlesinger and Ramankutty, 1994; Andronova and Schlesinger, 2000; Delworth and Mann, 2000); Knight et al. (2005) estimate that variations in the Atlantic Multidecadal Oscillation could account for up to 0.2°C peak-to-trough variability in Northern Hemisphere mean decadal temperatures.

9.4.1.3 Variability of Temperature from Observations and Models

Year to year variability of global mean temperatures simulated by the most recent models compares reasonably well with that of observations, as can be seen by comparing observed and modelled variations in Figure 9.5a. A more quantitative evaluation of modelled variability can be carried out by comparing the power spectra of observed and modelled global mean temperatures. Figure 9.7 compares the power spectrum of observations with those of transient simulations of the instrumental period. This avoids the need to compare variability estimated from long control runs of models with observed variability, which is difficult because observations are likely to contain a response to external forcings that cannot be reliably removed by subtracting a simple linear trend. The simulations considered contain both anthropogenic and natural forcings, and include most 20C3M simulations in the multi-model data archive at PCMDI (MMD). Figure 9.7 shows that the models have variance on global scales that is consistent with the observed variance at the 5% significance level on the decadal to inter-decadal time-scales important for detection and attribution. Figure 9.8 shows that this is also generally the case on continental scales, although model uncertainty is larger on smaller scales (Section 9.4.2.2).

[INSERT FIGURE 9.7 HERE]

[INSERT FIGURE 9.8 HERE]

Detection and attribution studies routinely assess if the residual variability unexplained by forcing is consistent with the estimate of internal variability (Allen and Tett, 1999; Tett et al., 1999; Stott et al., 2001; Zwiers and Zhang, 2003, etc). Furthermore, there is no evidence that the variability in paleoclimatic reconstructions that is not explained by forcing is stronger than that in models, and simulations of the last 1000 yrs show similar variability to reconstructions (Section 9.3.3.2). Chapter 8 discusses the simulation of major modes of variability in models and the extent to which they are simulated by models (including on decadal to inter-decadal time scales).

9.4.1.4 The Influence of Greenhouse Gas and Total Anthropogenic Forcing on Global Surface Temperature

Since the TAR, a large number of studies based on the longer observational record, improved models, and strengthening signal-to-noise ratio have increased confidence in the detection of an anthropogenic signal in the instrumental record (see, for example, the recent review by IDAG, 2005). Many more detection and attribution studies are now available than were available for the TAR, and these have used more recent climate data than previous studies and a much greater variety of climate simulations with more sophisticated treatments of a greater number of forcings of both anthropogenic and natural origins.

Fingerprint studies that use climate change signals estimated from an array of climate models indicate that detection of an anthropogenic contribution to the observed warming is a result that is robust to a wide range of model uncertainty, forcing uncertainties, and analysis techniques (Hegerl et al., 2001; Gillett et al., 2002c;

Tett et al., 2002; Zwiers and Zhang, 2003; IDAG, 2005; Stone and Allen, 2005b; Stone et al., 2006b; Stone et al., 2006a; Stott et al., 2006b; Stott et al., 2006c; Zhang et al., 2006). These studies account for the possibility that the agreement between simulated and observed global mean temperature changes could be fortuitous as a result of, for example, balancing too great (or too small) a model sensitivity by a too large (or too small) negative aerosol forcing (Schwartz, 2004; Hansen et al., 2005) or a too small (or too large) warming due to solar changes. Multi-signal detection and attribution analyses do not rely on such agreement because they seek to explain the observed temperature changes in terms of the responses to individual forcings, using model-derived patterns of response and a noise-reducing metric (Appendix 9.A), but determining their amplitudes from observations. As discussed in Section 9.2.2.1, these approaches make use of differences in the temporal and spatial response to forcings to separate their effect in observations.

Since the TAR there has also been an increased emphasis on quantifying the greenhouse gas contribution to observed warming, and distinguishing this contribution from other factors, both anthropogenic such as the cooling effects of aerosols and natural, such as from volcanic eruptions and changes in solar radiation
A comparison of results using four different models (Figure 9.9) shows that there is a robust identification of a significant greenhouse warming contribution to observed warming that is likely greater than the observed warming over the last 50 years with a significant net cooling from other anthropogenic forcings over that period, dominated by aerosols. Stott et al. (2006c) compared results over the 20th century obtained using the HadCM3, PCM and GFDL R30 models. They found consistent estimates for the greenhouse gas attributable warming over the century, expressed as the difference between temperatures in the last and first decades of the century of 0.6 to 1.3°C (5 to 95%) offset by cooling from other anthropogenic factors associated mainly with cooling from aerosols of 0.1 to 0.7°C and a small net contribution from natural factors over the century of –0.1 to 0.1°C (Figure 9.9b). Scaling factors for the model response to three forcings are shown in Figure 9.9a. A similar analysis for the MIROC model finds a somewhat larger warming contribution from greenhouse gases of 1.2 to 1.5°C offset by a cooling of 0.6 to 0.8°C from other anthropogenic factors and a very small net natural contribution (Figure 9.9). In all cases, the 5th percentile of the warming attributable to greenhouse gases is greater than the observed warming over the last 50 years of the 20th century (Figure 9.9, bottom panels).

The detection and estimation of a greenhouse gas signal is also robust to more fully accounting for model uncertainty. An analysis which combines results from three climate models and thereby incorporates uncertainty in the response of these three models (by including an estimate of the inter-model covariance structure in the regression method; Huntingford et al., 2006), supports the results from each of the models separately that it is likely that greenhouse gases would have caused more warming than was observed over the 1950-1999 period (Figure 9.9, results labelled "EIV"). These results are consistent with the results of an earlier analysis, which calculated the mean response patterns from five models and included a simpler estimate of model uncertainty (obtained by a simple rescaling of the variability estimated from a long control run, thereby assuming that inter-model uncertainty has the same covariance structure as internal variability; Gillett et al., 2002c). Both the results of Gillet et al. (2002c) and Huntingford et al. (2006) indicate that inter-model differences do not greatly increase detection and attribution uncertainties and that averaging fingerprints improves detection results.

[INSERT FIGURE 9.9 HERE]

A robust anthropogenic signal is also found in a wide range of climate models that do not have the full range of simulations required to directly estimate the responses to individual forcings that would be required for the full multi-signal detection and attribution anayses (Stone et al., 2006b; Stone et al., 2006a). In these cases an estimate of the model's pattern of response to each individual forcing can be diagnosed by fitting a series of energy balance models, one for each forcing, to the mean coupled model response from all the forcings to diagnose the time-dependent response in the global mean for each individual forcing. The magnitude of these time-only signals can then be inferred from observations using detection methods (Stone et al., 2006b; Stone et al., 2006a). When applied to 13 different climate models which had transient simulations of 1901-2005 temperature change, Stone et al. (2006a) found that there was a robust detection across the models of greenhouse gas warming over this period, although uncertainties in attributable temperature changes due to the different forcings were larger than when considering spatio-temporal patterns. By tuning an EBM to the observations, and using an AOGCM solely to estimate internal variability, Stone and Allen (2005b) detected the effects of greenhouse gases and tropospheric sulphate aerosols in the observed 1900-2004 record, but not the effects of volcanic and solar forcing.

The detection of an anthropogenic signal is also robust to using different methods. For example, Bayesian detection analyses (Appendix 9.A.2) robustly detect anthropogenic influence on near-surface temperature changes (Smith et al., 2003; Schnur and Hasselmann, 2005; Min and Hense, 2006a, 2006b). In these studies, Bayes Factors (ratios of posterior to prior odds) are used to assess evidence supporting competing hypotheses (Kass and Raftery, 1995; see appendix 9.A.2). A Bayesian analysis of seven climate models (Schnur and Hasselman (2005) and Bayesian analyses of MMD 20C3M simulations (Min and Hense, 2006a, 2006b)) found decisive evidence for the influence of anthropogenic forcings. Lee et al. (2005) using an approach suggested by Berliner et al. (2000) evaluated the evidence for the presence of the combined greenhouse gas and sulphate aerosols (GS) signal, estimated from CGCM1 and CGCM2 (Table 8.1, McAvaney et al., 2001), in observations for several 5-decade windows, beginning with 1900–1949, and ending with 1950–1999. Very strong evidence was found in support of detection of the forced response during both halves of the 20th century regardless of the choice of prior distribution. However, evidence for attribution in that approach is based on the extent to which observed data narrow the prior uncertainty on the size of the anthropogenic signal. That evidence was not found to be very strong, although Lee et al. (2005) estimate that strong evidence for attribution as defined in their approach may emerge within the next two decades as the anthropogenic signal strengthens.

In a further study, Lee et al. (2006) assess whether anthropogenic forcing has enhanced the predictability of decadal global scale temperature changes; a forcing related enhancement in predictability would give a further indication of its role in the evolution of the 20th century climate. Using an ensemble of simulations of the 20th century with GS forcing, they use Bayesian tools similar to those of Lee et al. (2005) to produce, for each decade beginning with 1930–1939, a forecast of the probability of above normal temperatures where "normal" is defined as the mean temperature of the preceding three decades. These hindcasts become skilful during the last two decades of the 20th century as indicated both by their Brier skill scores, a standard measure of the skill of probabilistic forecasts, and the confidence bounds on hindcasts of global mean temperature anomalies (Figure 9.10). This indicates that greenhouse gas forcing contributes to predictability of decadal temperature changes during the latter part of the 20th century.

[INSERT FIGURE 9.10 HERE]

Another type of analysis is a Granger causality analysis of the lagged covariance structure of observed hemispheric temperatures (Kaufmann and Stern, 2002), which also provides evidence for an anthropogenic signal, although such evidence may not be conclusive on its own without additional information from climate models (Triacca, 2001). Consistently, a neural network model is unable to reconstruct the observed global temperature record from 1860 to 2000 if anthropogenic forcings are not taken into account (Pasini et al., 2006). Further, an assessment of recent climate change relative to the long term persistence of Northern Hemisphere mean temperature as diagnosed from a range of reconstructed temperature records (Rybski et al., 2006) suggests that the recent warming cannot be explained solely in terms of natural factors, regardless of the reconstruction used. Similarly, Fomby and Vogelsang (2002), using a serial-correlation robust test of trend, found that the increase in global mean temperature over the 20th century is statistically significant even if it is assumed that natural climate variability has strong serial correlation.

9.4.1.5 The Influence of Other Anthropogenic and Natural Forcings

A significant cooling from other anthropogenic factors, dominated by aerosols, is a robust feature of a wide range of detection analyses. These analyses indicate that it is likely that greenhouse gases alone would have caused more than the observed warming over the last 50 years of the 20th century, with some warming offset by cooling from natural and other anthropogenic factors, notably aerosols, which have a very short residence time in the atmosphere relative to that of well mixed greenhouse gases (Schwartz, 1993). A key factor in identifying the aerosol fingerprint, and therefore the amount of aerosol cooling counteracting greenhouse warming, is the change through time of the hemispheric temperature contrast, which is affected by the different evolution of aerosol forcing in the two hemispheres as well as the greater thermal inertia of the larger ocean area in the Southern Hemisphere (Santer et al., 1996b; Santer et al., 1996c; Hegerl et al., 2001; Stott et al., 2006c). Regional and seasonal aspects of the temperature response may help to distinguish further the response to greenhouse gas increases from the response to aerosols (e.g., Ramanathan et al., 2005; Nagashima et al., 2006).

Results on the importance and contribution from anthropogenic forcings other than greenhouse gases vary more between different approaches. For example, Bayesian analyses differ in the strength of evidence they find for an aerosol effect. Schnur and Hasselman (2005), for example, failed to find decisive evidence for the influence of aerosols. They postulated that this could be due to taking account of modelling uncertainty in the response to aerosols. However, two other studies using frequentist methods that also include modelling uncertainty find a clear detection of sulphate aerosols, suggesting that the use of multiple models helps to reduce uncertainties and improves detection of a sulphate aerosol effect (Gillett et al., 2002c; Huntingford et al., 2006). Similarly, a Bayesian study of hemispheric mean temperatures from 1900-1996 finds decisive evidence for an aerosol cooling effect (Smith et al., 2003). Differences in separate detection of sulphate aerosol influences in a multi-signal approach can also reflect differences in the diagnostics applied (e.g., the space-time analysis of Tett et al. (1999) versus the space only analysis of Hegerl et al. (1997; 2000)) as was shown by Gillett et al. (2002a).

Recent estimates (Figure 9.9) indicate a relatively small combined effect of natural forcings on the global mean temperature evolution of the second half of the 20th century, with a small net cooling from the combined effects of solar and volcanic forcings. Coupled models simulate much less warming over the 20th century in response to solar forcing alone than to greenhouse gas forcing (Cubasch et al., 1997; Broccoli et al., 2003; Meehl et al., 2004), independent of which solar forcing reconstruction is used (Chapter 2). Several studies have attempted to estimate the individual contributions from solar and volcanic forcings separately, thus allowing for the possibility of enhancement of the solar response in observations due to processes not represented in models. Optimal detection studies that attempt to separate the responses to solar and other forcings in observations can also account for gross errors in the overall magnitude of past solar forcing, which remains uncertain (Chapter 2), by scaling the space-time patterns of response (Section 9.2.2.1). Using such a method, Tett et al. (1999) estimate that the net anthropogenic warming in the second half of the 20th century was much greater than any possible solar warming, even when using the solar forcing reconstruction by Hoyt and Schatten (1993), which indicates larger solar forcing, and a different evolution over time, than more recent reconstructions (Chapter 2, Section 2.7.1). However, Stott et al. (2003b), using the same solar reconstruction but a different model, were not able to completely rule out the possibility that solar forcing might have caused more warming than greenhouse gas forcing over the 20th century due to difficulties in distinguishing between the patterns of response to solar and greenhouse forcing. This was not the case when using the response to solar forcing based on the alternative reconstruction of Lean et al. (1995) in which case they found a very small likelihood (less than 1%, as opposed to approximately 10%) that solar warming could be greater than greenhouse warming since 1950. Note that recent solar forcing reconstructions show a substantially decreased magnitude of low-frequency variations in solar forcing (Chapter 2, Section 2.7.1) compared to Lean et al. (1995), and particularly, Hoyt and Schatten (1993).

The conclusion that greenhouse warming dominates over solar warming is supported further by a detection and attribution analysis using 13 models from the multi-model dataset (Stone et al., 2006a) and an analysis of the NCAR CCSM 1.4 model (Stone et al., 2006b). In both these analyses the response to solar forcing in the model was inferred by fitting a series of energy balance models to the mean coupled model response to the combined effects of anthropogenic and natural forcings. In addition, a combined analysis of the response at the surface and through the depth of the atmosphere using HadCM3 and the solar reconstruction of Lean et al. (1995) concluded that the near-surface temperature response to solar forcing over 1960–1999 is much smaller than the response to greenhouse gases (Jones et al., 2003). This conclusion is also supported by the vertical pattern of climate change, which is more consistent with the response to greenhouse gas than solar forcing (Figure 9.1). Further evidence against a dominant solar role arises from older analyses targeted at detecting the solar response (e.g., North and Stevens, 1998). Based on these detection results, which allow for possible amplification of the solar influence by processes not represented in climate models, we conclude that it is very likely that greenhouse gases caused more global warming over the last 50 years than changes in solar irradiance.

Detection and attribution as well as modelling studies indicate more uncertainty regarding the causes of early 20th century warming than the recent warming. A number of studies detect a significant natural contribution to early 20th century warming (Tett et al., 2002; Stott et al., 2003b; Nozawa et al., 2005; Shiogama et al., 2006). Some studies find a greater role for solar forcing than other forcings before 1950 (Stott et al., 2003b), although one detection study finds a roughly equal role for solar and volcanic forcing (Shiogama et al., 2006), whilst others find that volcanic forcing (Hegerl et al., 2003; Hegerl et al., 2006b) or a substantial contribution from natural internal variability (Tett et al., 2002; Hegerl et al., 2006b) could be important. There could also be an early expression of greenhouse warming in the early 20th century (Tett et al., 2002; Hegerl et al., 2003; Hegerl et al., 2006b).

9.4.1.6 Implications for Transient Climate Response

Quantification of the likely contributions of greenhouse gases and other forcing factors to past temperature change (Section 9.4.1.4) in turn provides observational constraints on the transient climate response, which determines the rapidity and strength of a global temperature response to external forcing (see Glossary, Section 9.6.2.3 and Chapter 8, Section 8.6.2.1 for detailed definitions) and therefore helps to constrain likely future rates of warming. Scaling factors derived from detection analyses can be used to scale predictions of future change by assuming that the fractional error in model predictions of global mean temperature change is constant (Allen et al., 2000; Allen et al., 2002; Allen and Stainforth, 2002; Stott and Kettleborough, 2002). This linear relationship between past and future fractional error in temperature change has been found to be sufficiently robust over a number of realistic forcing scenarios to introduce little additional uncertainty (Kettleborough et al., 2006). In this approach based on detection and attribution methods, which is compared with other approaches for producing probabilistic projections in Chapter 10, Section 10.5.4.5, different scalings are applied to the greenhouse gases and to the response to other anthropogenic forcings (notably aerosols); these separate scaling factors are used to account for possible errors in the models and aerosol forcing. Uncertainties calculated this way are likely to be more reliable than uncertainty ranges derived from simulations by coupled ocean-atmosphere general circulation models that happen to be available. Such ensembles could provide a misleading estimate of forecast uncertainty because they do not systematically explore modelling uncertainty (Allen et al., 2002; Allen and Stainforth, 2002). Stott et al. (2006c) compared observationally constrained predictions from three coupled climate models with a range of sensitivities and showed that predictions made in this way were relatively insensitive to the particular choice of model used to produce them. The robustness to choice of model of such observationally constrained predictions was also demonstrated by Stone et al. (2006a) for the multi-model dataset (MMD) ensemble. The observationally constrained transient climate response at the time of doubling of carbon dioxide following a 1% per year increase in carbon dioxide was estimated by Stott et al. (2006c) to lie between 1.5 and 2.8°C (Section 9.6.1, Figure 9.21). Such approaches have also been used to provide observationally constrained predictions of global mean (Stott and Kettleborough, 2002; Stone et al., 2006a) and continental scale temperatures (Stott et al., 2006a) following SRES emissions scenarios, and these are discussed in Chapter 10, Section 10.5.4.5 and Chapter 11, Section 11.10.

9.4.1.7 Studies of Indices for Temperature Change

Another method for identifying fingerprints of climate change in the observational record is to use simple indices of surface air temperature patterns that reflect features of the anticipated response to anthropogenic forcing (Karoly and Braganza, 2001; Braganza et al., 2003). By comparing modelled and observed changes in such indices, which included the global-mean surface temperature, the land-ocean temperature contrast, the temperature contrast between Northern and Southern Hemispheres, the mean magnitude of the annual cycle in temperature over land, and the mean meridional temperature gradient in the Northern Hemisphere mid-latitudes, Braganza et al. (2004) estimate that anthropogenic forcing accounts for almost all of the warming observed between 1946 and 1995 whereas warming between 1896 and 1945 was explained by a combination of anthropogenic and natural forcing and internal variability. These results are consistent with the results from studies using space-time detection techniques (Section 9.4.1.4).

Diurnal temperature range (DTR) has decreased over land by about 0.4°C over the last 50 years, with most of that change occurring prior to 1980 (Chapter 3, Section 3.2.2.1). This decreasing trend has been shown to be outside the range of natural internal variability estimated from models. Hansen et al. (1995) demonstrated that tropospheric aerosols plus increases in continental cloud cover, possibly associated with aerosols, could account for the observed decrease in DTR. However, although models simulate a decrease in DTR when they include anthropogenic changes in greenhouse gases and aerosols, the observed decrease is larger than the model-simulated decrease (Stone and Weaver, 2002, 2003; Braganza et al., 2004). This discrepancy is associated with simulated increases in daily maximum temperature being larger than observed, and could be associated with simulated increases in cloud cover being smaller than observed (Braganza et al., 2004; see Section 3.4.3.1 for observations), a result supported by other analyses (Dai et al., 1999; Stone and Weaver, 2002, 2003).

9.4.1.8 Remaining Uncertainties

A much larger range of forcing combinations and climate model simulations has been analysed in detection studies than were available for the TAR (Supplementary Material, Table S9.1), Detection and attribution analyses show robust evidence for an anthropogenic influence on climate. However, some forcings are still omitted by many models and uncertainties still remain in the treatment of those forcings that are included by the majority of models.

Most studies omit two forcings that could have significant effects, particularly on regional scales, namely carbonaceous aerosols, and land use changes. However, detection and attribution analyses based on climate simulations that include these forcings, (e.g., Stott et al., 2006b), continue to detect a significant anthropogenic influence in 20th century temperature observations even though the near-surface patterns of response to black carbon aerosols and sulphate aerosols could be so similar on large spatial scales (although opposite in sign) that detection analyses may be unable to distinguish between them (Jones et al., 2005). Forcing from surface albedo changes due to land use change is expected to be negative globally (Chapter 2, Section 2.5.3; Chapter 7, Section 7.3.3; Sections 9.2.1.1 and 9.3.3.3) although tropical deforestation could increase evaporation and warm the climate (Chapter 2, Section 2.5.5) counter-acting cooling from albedo change; however the albedo-induced cooling effect is expected to be small and was not detected in observed trends in the study by Matthews et al. (2004).

For those forcings that have been included in attribution analyses, uncertainties associated with the temporal and spatial pattern of the forcing and the modelled response can affect the results. Large uncertainties associated with estimates of past solar forcing (Chapter 2, Section 2.7.1) and omission of some chemical and dynamical response mechanisms (Gray et al., 2005) make it difficult to reliably estimate its contribution to warming over the 20th century. Nevertheless, as discussed above, results generally indicate that the contribution is small even if allowance is made for amplification of the response in observations, and simulations used in attribution analyses use several different estimates of solar forcing changes over the 20th century (Supplementary Material, Table S9.1). A number of different volcanic reconstructions are included in the modelling studies described in Section 9.4.1.2 (e.g., Sato et al., 1993; Andronova et al., 1999; Ammann et al., 2003; Supplementary Material, Table S9.1). Some models include volcanic effects by simply perturbing the incoming shortwave radiation at the top of the atmosphere, whilst others simulate explicitly the radiative effects of the aerosols in the stratosphere. Also, some models include the indirect effects of tropospheric sulphate aerosols on clouds (e.g., Tett et al., 2002), whereas others consider only the direct radiative effect (e.g., Meehl et al., 2004). In models that include indirect effects, different treatments of the indirect effect are used including changing the albedo of clouds according to an off-line calculation (e.g., Tett et al., 2002) and a fully interactive treatment of the effects of aerosols on clouds (e.g., Stott et al., 2006b). The overall level of consistency between attribution results derived from different models (as shown in Figure 9.9), and the ability of climate models to simulate large-scale temperature changes during the 20th century (Figures 9.5 and 9.6), indicate that such model differences are likely to have a relatively small impact on attribution results of large scale temperature change at the surface.

There have also been methodological developments that have resulted in attribution analyses taking uncertainties more fully into account. Attribution analyses normally directly account for errors in the magnitude of the model's pattern of response to different forcings by the inclusion of factors that scale the model responses up or down to best match observed climate changes. These scaling factors compensate for under- or over-estimates of the amplitude of the model response to forcing that may result from factors such as errors in the model's climate sensitivity, ocean heat uptake efficiency or errors in the imposed external forcing. Older analyses (e.g., Tett et al., 2002) did not take account of uncertainty due to sampling signal estimates from finite-member ensembles. This can lead to a low bias, particularly for weak forcings, in the scaling factor estimates (Appendix 9.A.1; Allen and Stott, 2003; Stott et al., 2003a). However, taking account of sampling uncertainty (as most more recent detection and attribution studies do, including those shown in Figure 9.9), makes relatively little difference to estimates of attributable warming rates, particularly to greenhouse gases, with the largest differences being to estimates of upper bounds for small signals, such as the response to solar forcing (Allen and Stott, 2003; Stott et al., 2003a). Studies that compare results between models and analysis techniques (e.g., Hegerl et al., 2000; Gillett et al., 2002a; Hegerl and Allen, 2002), and more recently, that use multiple models to determine fingerprints of climate change (Gillett et al., 2002c; Huntingford et al., 2006; Stott et al., 2006c; Zhang et al., 2006) find a robust detection of an anthropogenic signal in past temperature change.

A common aspect of detection analyses is that they assume the response in models to combinations of forcings to be additive. This was shown to be the case for near-surface temperatures in the PCM model (Meehl et al., 2004), in the HadCM2 model (Gillett et al., 2004c) and in the GFDL CM2.1 model (Knutson et al., 2006) although none of these studies considered the indirect effects of sulphate aerosols. Sexton et al. (2003) did find some evidence for a non-linear interaction between the effects of greenhouse gases and the indirect effect of sulphate aerosols in the atmosphere only version of HadCM3 forced by observed SSTs; the additional effect of combining greenhouse gases and indirect aerosol effects together was much smaller than each term separately but was found to be comparable to the warming due to increasing tropospheric ozone. In addition, Meehl et al. (2003) found that additivitiy does not hold so well for regional responses to solar and greenhouse forcing in the PCM. Linear additivity was found to hold in the PCM model for changes in tropopause height and synthetic satellite-borne microwave sounder (MSU) temperatures (Christy et al., 2000; Mears et al., 2003; Santer et al., 2003b).

A further source of uncertainty derives from the estimates of internal variability that are required for all detection analyses. These estimates are generally model-based because of difficulties in obtaining reliable internal variability estimates from the observational record on the space and time scales considered in detection studies. However, models would need to underestimate variability by factors of over 2 in their standard deviation to nullify detection of greenhouse gases in near surface temperature data (Tett et al., 2002) which appears unlikely given the quality of agreement between models and observations on global and continental scales (Figures 9.7, 9.8) and agreement with inferences on temperature variability from Northern Hemisphere temperature reconstructions of the last millennium. The detection of the effects of other forcings, including aerosols, is likely to be more sensitive (e.g., an increase of 40% in the estimate of internal variability is enough to nullify detection of aerosol and natural forcings in HadCM3; Tett et al., 2002)

Few detection studies have explicitly considered the influence of observational uncertainty on near-surface temperature changes. However, Hegerl et al. (2001) showed that inclusion of observational sampling uncertainty had relatively little effect on detection results and that random instrumental error had even less effect. Systematic instrumental errors, such as changes in measurement practices or urbanization, could be more important, especially earlier in the record (Chapter 3), although these errors are calculated to be relatively small on large spatial scales. Urbanization effects appear to have negligible effects on continental and hemispheric average temperatures (Chapter 3). Observational uncertainties are likely to be more important for surface temperature changes averaged over small regions and also for analyses of free atmosphere temperature changes (Section 9.4.4).

9.4.2 Continental and Subcontinental Surface Temperature Change

9.4.2.1 Observed Changes

Over the 1901–2005 period there has been warming over most of the Earth's surface with the exception of an area south of Greenland and parts of North and South America (Chapter 3, Figure 3.9 and Section 3.2.2.7, see also Figure 9.6). Warming has been strongest over the continental interiors of Asia and northwestern North America and some mid-latitude ocean regions of the Southern Hemisphere as well as southeastern Brazil. Since 1979, almost all land areas with observational data coverage show warming (Figure 9.6). Warming is smaller in the Southern Hemisphere than the Northern Hemisphere with cooling over parts of the mid-latitude oceans. There have been widespread decreases in continental diurnal temperature range since the 1950s which coincide with increases in cloud amounts (Chapter 3, Section 3.4.3.1).

9.4.2.2 Studies Based on Space Time Patterns

Global-scale analyses using space-time detection techniques (Section 9.4.1.4) have robustly identified the influence of anthropogenic forcing on the 20th century global climate. A number of studies have now extended these analyses to also consider sub-global scales. Two approaches have been used; one to assess the extent to which global studies can provide information on sub-global scales, the other to assess the influence of external forcing on the climate in specific regions. Limitations and problems in using smaller spatial scales are discussed at the end of this section.

The approach taken by IDAG (2005) was to compare analyses of full space-time fields with results obtained after removing the globally averaged warming trend, or after removing the annual global mean from each year in the analysis. They found that the detection of anthropogenic climate change is driven by the pattern of the observed warming in space and time, not just by consistent global mean temperature trends between models and observations. These results suggest that greenhouse warming should also be detectable on sub-global scales (see also Barnett et al., 1999). IDAG (2005) also showed that uncertainties increase, as expected, when global mean information, which has a high signal-to-noise ratio, is disregarded (see also North et al., 1995).

Another approach for assessing the regional influence of external forcing is to apply detection and attribution analyses to observations in specific continental or sub-continental-scale regions. A number of studies using a range of models and examining various continental or sub-continental scale land areas find a detectable human influence on 20th century temperature changes, either by considering the 100-year period from 1900 or the 50-year period from 1950. The warming effects of increasing greenhouse gas concentrations were detected by Stott (2003) in six continental-scale regions over the 1900-2000 period, using HadCM3 simulations. In most regions, cooling from sulphate aerosols was found to counteract some of the greenhouse warming. However, the separate detection of a sulphate aerosol signal in regional analyses remains difficult because of lower signal-to-noise ratios, loss of large scale spatial features of response such as hemispheric asymmetry that helps to distinguish different signals, and greater modelling and forcing uncertainty on smaller scales. Human influence was also detected using the CCCma model by Zwiers and Zhang (2003) over the 1950-2000 period in a series of nested regions, beginning with the full global domain and descending to separate continental domains for North America and Eurasia. Zhang et al. (2006) update this study using additional models (HadCM2 and HadCM3). They find evidence that climates in both continental domains have been influenced by anthropogenic emissions during 1950-2000, and generally also in the sub-contental domains (Figure 9.11). This finding is robust to the exclusion of North Atlantic Oscillation (NAO/AO) related variability, which is associated with part of the warming in Central Asia and could itself be related to anthropogenic forcing (Section 9.5.3). As the spatial scales considered become smaller, the uncertainty in estimated signal amplitudes (as demonstrated by the size of the vertical bars in Figure 9.11) becomes larger, reducing the signal-to-noise ratio (see also Stott and Tett, 1998). The signal-to-noise ratio, however, also depends on the strength of the climate change and the local level of natural variability, and therefore differs between regions. Most of the results noted above hold even if the estimate of internal climate variability from the control simulation is doubled.

[INSERT FIGURE 9.11 HERE]

The ability of models to simulate many features of the observed temperature changes and variability on continental and sub-continental scales and the detection of anthropogenic effects on each of six continents provides stronger evidence of human influence on climate than was available to the TAR. A comparison between a large ensemble of 20th century simulations of regional temperature changes made with the multi-model dataset at PCMDI (using the same simulations for which the global mean temperatures are plotted in Figure 9.5) shows that the spread of the multi-model ensembles encompasses the observed changes in regional temperature changes in almost all sub-continental regions (Figure 9.12; see also FAQ 9.2, Figure 1, and related figures in Chapter 11). There is a clear separation of the ensembles of simulations including just natural forcings from the ensembles of simulations containing both anthropogenic and natural forcings in many of the regions. A more detailed analysis of one particular model, HadCM3, shows that it reproduces many features of the observed temperature changes and variability in the different regions (IDAG, 2005). The GFDL CM2 model is also able to reproduce many features of the evolution of temperature change in a number of regions of the globe (Knutson et al., 2006). Other studies show success at simulating regional temperatures when models include anthropogenic and natural forcings. Wang et al. (2006) showed that all MMD 20C3M simulations replicated the late 20th century Arctic warming to various degrees, while both forced simulations and control simulations reproduce multi-year Arctic warm anomalies similar in magnitude to the observed mid 20th-century warming event.

[INSERT FIGURE 9.12 HERE]

There is some evidence that an anthropogenic signal can now be detected in some sub-continental scale areas using formal detection methods (Appendix 9.A.1), although this evidence is weaker than on continental scales. Zhang et al. (2006) detect anthropogenic fingerprints in China and southern Canada. Spagnoli et al. (2002) find some evidence for a human influence on 30-year trends of summer minimum daily temperatures in France, but they use a fingerprint estimated from a simulation of future climate change and they do not detect an anthropogenic influence on the other indices they consider, including summer maximum daily temperatures and winter temperatures. An anthropogenic influence on East Asian temperature changes was found by Min et al. (2005) in a Bayesian framework, but they did not consider anthropogenic aerosols or natural forcings in their analysis. Atmosphere-only general circulation model simulations forced with observed sea surface temperatures can potentially detect anthropogenic influence on smaller spatial and temporal scales than coupled model analyses but have the weakness that they do not explain the observed SST changes (Sexton et al., 2003). SST changes (Sexton et al., 2003). Two studies have applied attribution analysis on sub-continental temperatures to make inferences about changes in related variables. Stott et al. (2004) detect an anthropogenic influence on southern European summer mean temperature changes of the past 50 years and then infer the likelihood of exceeding an extreme temperature threshold (Section 9.4.3.3). Gillett et al. (2004a) detect an anthropogenic contribution to summer season warming in Canada and demonstrate a statistical link with area burned in forest fires. However, the robustness of these results to factors such as the choice of model or analysis method remains to be established given the limited number of studies on sub-continental scales.

Knutson et al. (2006) assessed temperature changes in regions of the world covering between 0.3% to 7.4% of the area of the globe and including tropical and extratropical land and ocean regions. They found much better agreement between climate simulations and observations when the models included rather than excluded anthropogenic forcings, which suggests a detectable anthropogenic warming signal over many of the regions they examined. This would indicate the potential for formal detection studies to detect anthropogenic warming in many of these regions, although Knutson et al. (2006) also note that in some regions the climate simulations they examined were not very realistic and showed that some of these discrepancies are associated with modes of variability such as the Arctic Oscillation.

Karoly and Wu (2005) compared observed temperature trends in 5º x 5º grid boxes globally over 30, 50 and 100-yr periods ending in 2002 with 1) internal variability as simulated by three models (GFDL-R30, HadCM2, PCM), and 2) the simulated response to greenhouse gas and sulphate aerosol forcing in those models (see also, Knutson et al., 1999). They find that a much higher percentage of gridboxes show trends that are inconsistent with model estimated internal variability than would be expected by chance and that a large fraction of gridboxes show changes that are consistent with the forced simulations, particularly over the two shorter periods. This assessment is essentially a global-scale detection result because its interpretation relies upon a global composite of grid-box scale statistics. As discussed in the paper, this result does not rule out the possibility that individual gridbox trends may be explained by different external forcing combinations, particularly since natural forcings and also forcings that could be important on small spatial scales, such as land use change or black carbon aerosols, are missing from these models. The demonstration of local consistency between models and observations in this study does not necessarily imply that observed changes can be attributed to anthropogenic forcing in a specific gridbox, and it does not allow confident estimates of the anthropogenic contribution to change at those scales.

Models do not reproduce the observed temperature changes equally well in all regions. Areas where temperature changes are not particularly well simulated by some models include parts of North America (Knutson et al., 2006) and mid Asia (IDAG, 2005; Figure 9.12). This could be due to a regional trend or variation that was caused by internal variability (a result that models would not be expected to reproduce), uncertain forcings that are locally important, or model errors. Examples of uncertain forcings that play a small role globally, but could become more important regionally are the effects of land use changes (Sections 9.2, 9.3) or atmospheric brown clouds. The latter could be important in explaining observed temperature trends in South Asia and the Northern Indian Ocean (Ramanathan et al., 2005; see Chapter 2).

An analysis of the MMD 20C3M experiments indicates that multi-decadal internal variability could be responsible for some of the rapid warming seen in the central United States between 1901 and 1940 and rapid cooling between 1940 and 1979 (Kunkel et al., 2006). Also, regional temperature is more strongly influenced by variability and changes in climate dynamics, such as temperature changes associated with the NAO, which may itself show an anthropogenic influence (Section 9.5.3.2), or the AMO, which could in some regions and seasons be poorly simulated by models and could be confounded with the expected temperature response to external forcings. Thus the anthropogenic signal is likely to be more easy to identify in some regions than in others, with temperature changes in those regions most affected by multi-decadal scale variability being the most difficult to attribute, even if those changes are inconsistent with model estimated internal variability and therefore detectable.

The extent to which temperature changes on sub-continental scales can be attributed to anthropogenic forcings, and the extent to which it is possible to estimate the contribution of greenhouse gas forcing to regional temperature trends, remains a topic for further research. Idealised studies (e.g., Stott and Tett, 1998) suggest that surface temperature changes are detectable mainly on large spatial scales of the order of several thousand kilometres (although they also showed that as the signal of climate change strengthens in the 21st century, surface temperature changes are expected to become detectable at increasingly smaller scales). Robust detection and attribution is inhibited at the gridbox scales because it becomes difficult to separate the effects of the relatively well understood large-scale external influences on climate, such as greenhouse gas, aerosols, solar and volcanic forcing, from each other and from local influences that may not be related to these large scale forcings. This occurs because the contribution from internal climate variability increases on smaller scales, because the spatial details that can help to distinguish between different forcings at large scales are not available or unreliable at smaller scales, and because forcings that could be important on small spatial scales, such as land use change or black carbon aerosols, are uncertain and may not have been included in the models used for detection. Although models do not typically underestimate natural internal variability of temperature at continental scales over land (Figure 9.8), even at a grid box scale (Karoly and Wu, 2005), the credibility of small-scale details of climate simulated by models is lower than for large-scale features. While the large scale coherence of temperatures means that temperatures at a particular grid box should adequately represent a substantial part of the variability of temperatures averaged over the area of that grid box, the remaining variability from local scale processes and the upward cascades from smaller to larger scales via non-linear interactions may not be well represented in models at the grid box scale. Similarly, the analysis of shorter temporal scales also decreases the signal-to-noise ratio and the ability to use time information to distinguish between different forcings. This is why most detection and attributuion studies use temporal scales of 50 or more years.

9.4.2.3 Studies Based on Indices for Temperature Change and Temperature-Precipitation Relationships

Studies based on indices for temperature change support the robust detection of human influence on continental-scale land areas. Observed trends in indices of North American continental scale temperature change, (including the regional mean, the mean land-ocean temperature contrast, and the annual cycle) were found by Karoly et al. (2003) to be generally consistent with simulated trends under historical forcing from greenhouse gases and sulphate aerosols during the second half of the 20th century. In contrast they found only a small likelihood of agreement with trends driven by natural forcing only during this period. An analysis of changes in Australian mean, daily maximum and daily minimum temperatures and diurnal temperature range using 6 coupled climate models showed that it is likely that there has been a significant contribution to observed warming in Australia from increasing greenhouse gases and sulphate aerosols (Karoly and Braganza, 2005a). An anomalous warming has been found over all Australia (Nicholls, 2003) and in New South Wales (Nicholls et al., 2005) since the early 1970s, associated with a changed relationship between annual mean maximum temperature and rainfall. Whereas interannual rainfall and temperature variations are strongly inversely correlated, in recent decades temperatures have tended to be higher for a given rainfall than in previous decades. By removing the rainfall-related component of Australian temperature variations, thereby enhancing the signal-to-noise ratio, Karoly and Braganza (2005b) detected an anthropogenic warming signal in south eastern Australia, although their results are affected by some uncertainty associated with their removal of rainfall-related temperature variability. A similar technique applied to the Sudan and Sahel region improved the agreement between model simulations and observations of temperature change over the last 60 years in this region (Douville, 2006) and could improve the detectability of regional temperature signals over other regions where precipitation is likely to affect the surface energy budget (Trenberth and Shea, 2005).

9.4.3 Surface Temperature Extremes

9.4.3.1 Observed Changes

Observed changes in temperature extremes are consistent with the observed warming of the climate (Alexander et al., 2006) and are summarised in Chapter 3, Section 3.8.2.1. There has been a widespread reduction in the number of frost days in mid-latitude regions in recent decades, an increase in the number of warm extremes, particularly warm nights, and a reduction in the number of cold extremes, particularly cold nights. A number of regional studies all show patterns of changes in extremes consistent with a general warming, although the observed changes in the tails of the temperature distributions are generally not consistent with a simple shift of the entire distribution alone.

9.4.3.2 Global Assessments

Evidence for observed changes in short duration extremes generally depends on the region considered and the analysis method (IPCC, 2001). Global analyses have been restricted by the limited availability of quality controlled and homogenized daily station data. Indices of temperature extremes have been calculated from station data, including some indices from regions where daily station data are not released (Frich et al., 2002; Klein Tank and Können, 2003; Alexander et al., 2006). Kiktev et al. (2003) analysed a subset of such indices by using fingerprints from atmospheric model simulations driven by prescribed sea surface temperatures. They find significant decreases in the number of frost days and increases in the number of very warm nights over much of the Northern Hemisphere. Comparisons of observed and modelled trend estimates indicate that inclusion of anthropogenic effects in the model integrations improves the simulation of these changing temperature extremes, indicating that human influences are probably an important contributor to changes in the number of frost days and warm nights. Tebaldi et al. (2006) found that changes simulated by eight MMD models agreed well with observed trends in heat waves, warm nights and frost days over the last four decades.

Christidis et al. (2005) analysed a new gridded dataset of daily temperature data (Caesar et al., 2006) using the indices shown by Hegerl et al. (2004) to have a potential for attribution, namely the average temperature of the most extreme 1, 5, 10 and 30 days of the year. Christidis et al. (2005) detected robust anthropogenic changes in indices of extremely warm nights using signals estimated with the HadCM3 model, although with some indications that the model over-estimates the observed warming of warm nights. Human influence on cold days and nights was also detected but in this case the model underestimated the observed changes, significantly so in the case of the coldest day of the year. Anthropogenic influence was not detected in observed changes in extremely warm days.

9.4.3.3 Attributable Changes in the Risk of Extremes

Many important impacts of climate change may manifest themselves through a change in the frequency or likelihood of occurrence of extreme events. While individual extreme events cannot be attributed to external,influences, a change in the probability of such events might be attributable to external influences (Palmer, 1999; Palmer and Räisänen, 2002). One study has shown that anthropogenic forcings have significantly increased the risk of extremely warm summer conditions over Southern Europe, as was observed during the 2003 European heatwave. Stott et al. (2004) applied a methodology for making quantitative statements about change in the likelihood of such specific types of climatic events (Allen, 2003; Stone and Allen, 2005a), by expressing the contribution of external forcing to the risk of an event exceeding a specific magnitude. If P_1 is the probability of a climatic event (such as a heat wave) occurring in the presence of anthropogenic forcing of the climate system, and P_0 is the probability of it occurring if anthropogenic forcing had not been present, then the fraction of the current risk that is attributable to past greenhouse gas emissions is given by $FAR = 1 - P_0/P_1$ (Allen, 2003). Stott et al. (2004) applied the FAR concept to mean summer temperatures of a large part of continental Europe and the Mediterranean. Using a detection and attribution analysis, they determined that regional summer mean temperature had likely increased due to anthropogenic forcing, and that the observed change was inconsistent with natural forcing. They then used the HadCM3 model to estimate the FAR associated with a particular extreme threshold of regional summer mean temperature that was exceeded in 2003, but in no other year since the beginning of the record in 1851. Stott et al. (2004) estimate that it is very likely that human influence has more than doubled the risk of the regional summer mean temperature exceeding this threshold (Figure 9.13).

[INSERT FIGURE 9.13 HERE]

This study considered only continental-mean seasonal-mean temperatures. Consideration of shorter term and smaller scale heatwaves will require higher resolution modelling and will need to take complexities such as land surface processes into account (Schär and Jendritzky, 2004). Also, Stott et al. (2004) assumed no change in internal variability in the region they considered (which was the case in HadCM3 21st climate projections for summer mean temperatures in the region they considered) thereby ascribing the increase in risk only to an increase in mean temperatures (i.e., as shown in Technical Summary, Box 3.4, Figure 1, which illustrates how a shift in the mean of a distribution can cause a large increase in the frequency of extremes). However, there is some evidence for a weak increase in European temperature variability in summer (and a decrease in winter) for the period 1961–2004 (Scherrer et al., 2005), which could contribute to an increase in the likelihood of extremes. Schär et al. (2004) showed that the central European heat wave of 2003 could also be consistent with model predicted increases in temperature variability due to soil moisture and vegetation feedbacks. In addition, multi-decadal scale variability, associated with basin-scale changes in the Atlantic Ocean related to the Meridional Overturning Circulation (MOC) could have contributed to changes in European summer temperatures (Sutton and Hodson, 2005), although Klein Tank et al. (2005) showed evidence that patterns of change in European temperature variance in spring and summer are not consistent with patterns of change in temperature variance expected from natural variability. Meteorological aspects of the summer 2003 European heat wave are discussed in Chapter 3, Box 3.6.5.

9.4.4 Free Atmosphere Temperature

9.4.4.1 Observed Changes

Observed free atmosphere temperature changes are discussed in Chapter 3, Section 3.4.1 and a comprehensive review has been made by Karl et al. (2006). Radiosonde based observations (near globally since 1958) and satellite based temperature measurements (beginning in late 1978) show warming trends in the troposphere and cooling trends in the stratosphere. All datasets show that the global-mean and tropical troposphere has warmed from 1958 to the present, with the warming trend in the troposphere being slightly more than at the surface. Since 1979, it is likely that there is slightly greater warming in the troposphere than at the surface, although uncertainties remain in observational warming trends of tropospheric temperature and whether these are greater than or less than the surface trend. The range (due to different datasets) of the global mean tropospheric temperature trend since 1979 is 0.12 to 0.19°C/decade based on satellite based estimates (Chapter 3) compared to a range of 0.16 to 0.18°C/decade for the global surface warming. Whilst all datasets show that the stratosphere has cooled considerably from 1958 and from 1979 to present, there are large differences in the linear trends estimated from different datasets. However, a linear trend is a poor fit to the data in the stratosphere and the tropics at all levels (Chapter 3, Section 3.4.1). The uncertainties in the observational records are discussed in detail in Chapter 3, Section 3.4.1 and by Karl et al. (2006). Uncertainties remain in homogenized radiosonde datasets which could result in a spurious inference of net cooling in the tropical troposphere. Differences between temperature trends measured from different versions of tropospheric satellite data result primarily from differences in how data from different satellites are merged.

9.4.4.2 Changes in Tropopause Height

The height of the lapse rate tropopause (the boundary between the stratosphere and the troposphere) is sensitive to bulk changes in the thermal structure of the stratosphere and the troposphere, and may also be affected by changes in surface temperature gradients (Schneider, 2004). Analyses of radiosonde data have documented increases in tropopause height over the past 3–4 decades (Highwood et al., 2000; Seidel et al., 2001). Similar increases have been inferred from three different reanalysis products (ERA-15, ERA-40 and NCAR-NCEP) (Kalnay et al., 1996; Gibson et al., 1997; Simmons and Gibson, 2000; Kistler et al., 2001) and from model simulations with combined anthropogenic and natural forcing (Santer et al., 2003b; Santer et al., 2003a; Santer et al., 2004; see Figure 9.14). In both models and reanalyses, changes in tropopause height over the satellite and radiosonde eras are smallest in the tropics and largest over Antarctica (Santer et al., 2003b; Santer et al., 2003a; Santer et al., 2004). Model simulations with individual forcings indicate that the major drivers of the model tropopause height increases are ozone-induced stratospheric cooling and the tropospheric warming caused by greenhouse gas increases (Santer et al., 2003a). However, earlier model studies have found that it is difficult to alter tropopause height through stratospheric ozone changes alone (Thuburn and Craig, 2000). Santer et al. (2003c) found that the model-simulated response to combined anthropogenic and natural forcing is robustly detectable in different reanalysis products, and that solar and volcanic forcing alone could not explain the tropopause height increases (Figure 9.14). Climate data from reanalyses, especially the "first generation" reanalysis analysed by Santer et al (2003a), are subject to some deficiencies, notably inhomogeneities related to changes over time in the availability and quality of input data to the reanalyses, and are subject to a number of specific technical choices in the reanalysis scheme (see Santer et al., 2004 for a discussion). Also, the NCEP reanalysis detection results could be due to compensating errors due to NCEPs excessive stratospheric cooling (Santer et al., 2004) since NCEP cools the troposphere more than models, while models tend to warm the troposphere more than NCEP. Differently, the finding of a significant anthropogenic influence on tropopause height in "second generation" ERA-40 reanalysis is driven by similar large-scale changes in models and that reanalysis. Detection results there are robust to removing global mean tropopause height increases.

[INSERT FIGURE 9.14 HERE]

9.4.4.3 Overall Atmospheric Temperature Change

Anthropogenic influence on free atmosphere temperatures has been detected in analyses of satellite data since 1979, although this finding has been found to be sensitive to which analysis of satellite data is used. Satellite-borne microwave sounders (MSU), beginning in 1978, estimate the temperature of thick layers of the atmosphere. The main layers represent the lower troposphere (T2LT), the mid-troposphere (T2) and the lower stratosphere (T4) (Chapter 3, Section 3.4.1.2.1). Santer et al. (2003c) compared T2 and T4 temperature changes simulated by the PCM model including anthropogenic and natural forcings with the UAH (Christy et al., 2000) and RSS (Mears and Wentz, 2005) satellite datasets (Chapter 3, Section 3.4.1.2.2). They found that the model fingerprint of the T4 response to combined anthropogenic and natural forcing was consistently detected in both satellite datasets, whereas the T2 response was detected only in the RSS dataset. However, when the global mean changes were removed, the T2 fingerprint was detected in both datasets, suggesting a common spatial pattern of response overlain by a systematic global mean difference.

Anthropogenic influence on free atmosphere temperatures has been robustly detected in a number of different studies analysing various versions of the HadRT2 radiosonde dataset (Parker et al., 1997) by means of a variety of different diagnostics and fingerprints estimated with the HadCM2 and HadCM3 models (Tett et al., 2002; Thorne et al., 2002; Jones et al., 2003; Thorne et al., 2003). Whereas an analysis of spatial patterns of zonal mean free atmosphere temperature changes was unable to detect the response to natural forcings (Tett et al., 2002), an analysis of spatio-temporal patterns detected the influence of volcanic aerosols, and (less convincingly) solar irradiance changes, in addition to detecting the effects of greenhouse gases and sulphate aerosols (Jones et al., 2003). Also, Crooks (2004) detected a solar signal in atmospheric temperature changes as seen in the HadRT2.1s radiosonde dataset when a diagnostic chosen to extract the solar signal from other signals was used. The models used in these studies have poor vertical resolution in the stratosphere and they significantly underestimate stratospheric variability, thus possibly overestimating the significance of these detected signals (Tett et al., 2002). However, a sensitivity study (Thorne et al., 2002) showed that detection of human influence on free atmosphere temperature changes does not depend on the inclusion of stratospheric temperatures. An analysis of spatial patterns of temperature change, represented by large scale area averages at the surface, in broad atmospheric layers and in lapse rates between layers, showed robust detection of an anthropogenic influence on climate when a range of uncertainties were explored relating to the choice of fingerprints and the radiosonde and model datasets (Thorne et al., 2003). However they were not able to attribute recent observed tropospheric temperature changes to any particular combination of external forcing influences because the models analysed (HadCM2 and HadCM3) over-estimate free-atmosphere warming as estimated by the radiosonde datasets, an effect also seen by Douglass et al. (2004) during the satellite era. However, there is evidence that radiosonde data during the satellite era are contaminated by spurious cooling trends (Sherwood et al., 2005; Randel and Wu, 2006; Chapter 3, Section 3.4.1), and since structural uncertainty arising from the choice of techniques used to analyse radiosonde data has not yet been quantified (Thorne et al., 2005), it is difficult to assess, based on these analyses alone, whether model-data discrepancies are due to model or observational deficiencies. However further information is provided by an analysis of modelled and observed tropospheric lapse rates which is discussed in Section 9.4.4.4.

A different approach is to assess detectability of observed temperature changes through the depth of the atmosphere with atmosphere-only general circulation model simulations forced with observed sea surface temperatures, although the vertical profile of the atmospheric temperature-change signal estimated in this way can be quite different from the same signal estimated by coupled models with the same external forcings (Hansen et al., 2002; Sun and Hansen, 2003; Santer et al., 2005). Sexton et al. (2001) found that inclusion of anthropogenic effects improved the simulation of zonally averaged upper air temperature changes from the HadRTt1.2 dataset such that an anthropogenic signal was detected at the 5% significance level in patterns of seasonal mean temperature change calculated as overlapping 8-year means over the 1976–1994 period and expressed as anomalies relative to the 1961–1975 base period. In addition, analysing patterns of annual mean mean temperature change for individual years showed that an anthropogenic signal was also detected on inter-annual time scales for a number of years toward the end of the analysis period.

9.4.4.4 Differential Temperature Trends

Subtracting temperature trends at the surface from those in the free atmosphere removes much of the common variability between these layers and tests whether the model-predicted trends in tropospheric lapse rate are consistent with those observed from radiosondes and satellites (Karl et al., 2006). Since 1979, globally averaged modelled trends in topospheric lapse rates are consistent with those observed. However, this is not the case in the tropics, where most models have more warming aloft than at the surface while most observational estimates show more warming at the surface than in the troposphere (Karl et al., 2006). Karl et al. (2006) carried out a systematic review of this issue. There is greater consistency beween simulated and observed differential warming in the tropics in some satellite measurements of tropospheric temperature change, particularly when the effect of the cooling stratosphere on tropospheric retrievals is taken into account (Karl et al., 2006). External forcing other than greenhouse gas changes can also help to reconcile some of the differential warming, since both volcanic eruptions and stratospheric ozone depletion are expected to have cooled the troposphere by more than the surface over the last several decades (Santer et al., 2000; IPCC, 2001; Santer et al., 2001; Free and Angell, 2002; Karl et al., 2006). There are, however, uncertainties in quantifying the differential cooling caused by these forcings, both in models and observations, arising from uncertainties in the forcings and model response to the forcings. Differential effects of natural modes of variability, such as ENSO and the NAM, on observed surface and tropospheric temperatures, which arise from differences in the amplitudes and spatial expression of these modes at the surface and in the troposphere, make only minor contributions to the overall differences in observed surface and tropospheric warming rates (Santer et al., 2001; Hegerl and Wallace, 2002; Karl et al., 2006).

A systematic intercomparison between radiosonde based (RATPAC, Free et al., 2005, and HadAT, Thorne et al., 2005) and satellite based (RSS, UAH) observational estimates of tropical lapse rate trends with those simulated by 19 MMD models shows that on monthly and annual timescales, variations of temperature at the surface are amplified aloft in both models and observations by consistent amounts (Santer et al., 2005; Karl et al., 2006). It is only on longer timescales that disagreement between modelled and observed lapse rates arises (Hegerl and Wallace, 2002), i.e. on the timescales on which discrepancies would arise from inhomogeneities in the observational record. Only one observational dataset (RSS) was found to be consistent with the models on both short and long timescales. Whilst Vinnikov et al. (2006) have not produced a lower tropospheric retrieval, their estimate of the T2 temperature trend (Chapter 3, Figure 3.18) is consistent with model simulations (Karl et al., 2006). One possibility is that amplification effects are controlled by different physical mechanisms on short and long time scales, although a more probable explanation is that some observational records are contaminated by errors that affect their long term trends (Chapter 3, Section 3.4.1, Karl et al., 2006).

9.4.5 Summary

Since the TAR, the evidence has strengthened that human influence has increased global temperatures near the surface of the Earth. Every year since the publication of the TAR has been in the top ten warmest years in the instrumental global record of near-surface temperatures. Many climate models are now available which simulate global mean temperature changes that are consistent with those observed over the last century when they include the most important forcings of the climate system. The fact that no coupled model simulation so far has reproduced global temperature changes over the 20th century without anthropogenic forcing is strong evidence for the influence of humans on global climate. This conclusion is robust to variations in model formulation and uncertainties in forcings as far as they have been explored in the large multi-model ensemble now available (Figure 9.5).

Many studies have detected a human influence on near-surface temperature changes, applying a variety of statistical techniques and using many different climate simulations. Comparison with observations shows that the models used in these studies appear to have an adequate representation of internal variability on the decadal to inter-decadal time-scales important for detection (Figure 9.7). When evaluated in a Bayesian framework, very strong evidence is found for a human influence on global temperature change regardless of the choice of prior distribution.

Since the TAR there has been an increased emphasis on partitioning the observed warming into contributions from greenhouse gas increases and other anthropogenic and natural factors. These studies lead to the conclusion that greenhouse gas forcing has very likely been the dominant cause of the observed global warming over the last 50 years, and account for the possibility that the agreement between simulated and observed temperature changes could be reproduced by different combinations of external forcing. This is because, in addition to detecting the presence of model simulated spatio-temporal response patterns in observations, such analyses also require consistency between the model simulated and observational amplitudes of these patterns.

Detection and attribution analyses indicate that over the past century there has likely been a cooling influence from aerosols and natural forcings counteracting some of the warming influence of the increasing concentrations of greenhouse gases (Figure 9.9). Spatial information is required in addition to temporal information to reliably detect the influence of aerosols and distinguish them from the influence of increased greenhouse gases. In particular, aerosols are expected to cause differential warming and cooling rates between the Northern and Southern Hemispheres which change with time depending on the evolution of the aerosol forcing, and this spatio-temporal fingerprint can help to constrain the possible range of cooling from aerosols over the century. Despite continuing uncertainties in aerosol forcing and the climate response, it is likely that greenhouse gases alone would have caused more warming than observed during the last fifty years, with some warming offset by cooling from aerosols and other natural and anthropogenic factors. The overall evidence from studies using instrumental surface temperature and free atmospheric temperature data, along with evidence from analysis of temperature over the last few hundred years (Section 9.3.3.2) indicates that it is very unlikely that the contribution from solar forcing to the warming of the last 50 years was larger than that from greenhouse gas forcing.

An important development since the TAR has been the detection of an anthropogenic signal in surface temperature changes since 1950 on continental and sub-continental scale land areas. The ability of models to simulate many aspects of the temperature evolution on these scales (Figure 9.12) and the detection of significant anthropogenic effects on each of six continents provides stronger evidence of human influence on the global climate than was available to the TAR. Difficulties remain in attributing temperature changes on smaller than continental scales and over timescales of less than 50 years. Attribution at these scales has, with limited exceptions, not yet been established. Temperature changes associated with some modes of variability, and which could be wholly or partly naturally caused, are poorly simulated by models in some regions and seasons and could be confounded with the expected temperature response to external forcings. Averaging over smaller regions reduces the natural variability less than averaging over large regions, making it more difficult to distinguish changes expected from external forcing. Also, the small-scale details of external forcing and the response simulated by models are less credible than large-scale features. Overall, uncertainties in observed and model simulated climate variability and change at smaller spatial scales make it difficult at present to estimate the contribution of anthropogenic forcing to temperature changes on scales smaller than continental and timescales shorter than 50 years.

There is now some evidence that anthropogenic forcing has affected extreme temperatures. There has been a significant decrease in the frequency of frost days and an increase in the incidence of warm nights. A detection and attribution analysis has shown a significant human influence on patterns of changes in extremely warm nights and evidence for a human-induced warming of the coldest nights and days of the year. Many important impacts of climate change are likely to manifest themselves through an increase in the frequency of heat-waves in some regions and a decrease in the frequency of extremely cold events in others. Based on a single study, and assuming a model based estimate of temperature variability, past human influence may have more than doubled the risk of European mean summer temperatures as high as those recorded in 2003 (Figure 9.13).

Since the TAR further evidence has accumulated that there has been a significant anthropogenic influence on free atmosphere temperature since wide-spread measurements became available from radiosondes in the late 1950s. The influence of greenhouse gases on tropospheric temperatures has been detected as has the influence of stratospheric ozone depletion on stratospheric temperatures. The combination of a warming troposphere and a cooling stratosphere has likely led to an increase in the height of the tropopause and model-data comparisons show that greenhouse gases and stratospheric ozone changes are likely largely responsible (Figure 9.14).

Whereas, on monthly and annual timescales, variations of temperature in the tropics at the surface are amplified aloft in both the MMD simulations and observations by consistent amounts, on longer timescales, simulations of differential tropical warming rates between the surface and the free atmosphere are inconsistent with some observational records. One possible explanation for the discrepancies on multi-annual but not shorter timescales is that amplification effects are controlled by different physical mechanisms, but a more probable explanation is that some observational records are contaminated by errors that affect their long term trends.

9.5 Understanding of Change in Other Variables during the Industrial Era

The objective of this section is to assess large-scale climate change in variables other than air temperature, including ocean climate changes, atmospheric circulation changes, precipitation changes, cryosphere changes and sea-level change. This section draws heavily on Chapters 3, 4, 5 and 8. Where possible, it attempts to identify links between changes in different variables, such as those that associate some aspects of sea-surface temperature change with precipitation change. It also discusses the role of external forcing, drawing where possible on formal detection studies.

9.5.1 Ocean Climate Change

9.5.1.1 Ocean Heat Content Changes

Since the TAR there has been an accumulation of evidence for climate change within the ocean, both at regional and global scales (Chapter 5). The overall heat content in the world ocean is estimated to have increased by 14.2×10^{22} J during the period 1961–2003 (Chapter 5). This overall increase has been superimposed on strong interannual and interdecadal variations. The fact that the entire ocean, which is by far the system's largest heat reservoir (Levitus et al., 2005; see also Figure 5.4) gained heat during the latter half of the 20th century is consistent with there having been a net positive radiative forcing of the climate system. Late 20th century ocean heat content changes were at least one order of magnitude larger than the increase in energy content of any other component of the earth's ocean-atmosphere-cryosphere system (Figure 5.4; Levitus et al., 2001).

All analyses indicate a large anthropogenic component to the positive trend in global ocean heat content. Levitus et al. (2001) and Gregory et al (2004) analysed simulations of the GFDL R30 and HadCM3 models respectively and showed that climate simulations agree best with observed changes when the models include anthropogenic forcings from increasing greenhouse gas concentrations and sulphate aerosols. Gent and Danabasoglu (2004) showed that the observed trend could not be explained by natural internal variability as simulated by a long control run of the CCSM2 climate model. Barnett et al. (2001) and Reichert et al. (2002b) used detection analyses similar to those described in Section 9.4 to detect model simulated ocean climate change signals in the observed spatio-temporal patterns of ocean heat content across the ocean basins.

Barnett et al. (2005) extended previous detection and attribution analyses of ocean heat content changes to a basin by basin analysis of the temporal evolution of temperature changes in the upper 700 m of the ocean (see also Pierce et al., 2006). They report that whereas the observed change is not consistent with internal variability and the response to natural external forcing as simulated by two climate models (PCM and HadCM3), the simulated ocean warming due to anthropogenic factors (including well mixed greenhouse gases and sulphate aerosols) is consistent with the observed changes and reproduces many of the different responses seen in the individual ocean basins (Figure 9.15), indicating a human-induced warming of the world's oceans with a complex vertical and geographical structure that is simulated quite well by the two AOGCMs. Barnett et al. (2005) found that the earlier conclusions of Barnett et al. (2001) were not affected by the (Levitus et al., 2005) revisions to the Levitus et al. (2000) ocean heat content data.

[INSERT FIGURE 9.15 HERE]

In contrast, changes in solar forcing can potentially explain only a small fraction of the observationally based estimates of increase in ocean heat content (Crowley et al., 2003) and the cooling influence of natural (volcanic) and anthropogenic aerosols is expected to have slowed ocean warming over the last half century. Delworth et al. (2005) find a delay of several decades and a reduction in the magnitude of the warming of approximately two thirds in simulations with the GFDL CM2 model which included these forcings, compared to the response to increasing greenhouse gases alone, consistent with results based on an upwelling diffusion energy balance model (Crowley et al., 2003). Reductions in ocean heat content are found following volcanic eruptions in climate simulations (Church et al., 2005), including a persistant century timescale signal of ocean cooling at depth following the eruption of Krakatoa (Gleckler et al., 2006).

Although the heat uptake in the ocean cannot be explained without invoking anthropogenic forcing, there is some evidence that the models have overestimated how rapidly heat has penetrated below the ocean's mixed layer (Forest et al., 2006; also see Figure 9.15). In simulations that include natural forcings in addition to anthropogenic forcings, eight coupled climate models simulate heat uptake of 0.26 ± 0.06 W m^{-2} (plus and minus one standard deviation) for 1961-2003, whereas observations of ocean temperature changes indicate 0.21 ± 0.04 W m^{-2} (Chapter 5, Section 5.2.2.1). These could be consistent within their uncertainties but might indicate a tendency of climate models to overestimate ocean heat uptake.

In addition, the interannual to decadal variability seen in Levitus et al. (2000; 2005) (Chapter 5, Section 5.2.2) is is underestimated by models; Gregory et al. (2004) showed significant differences between observed and modeled interannual deviations from a linear trend in five-year running means for world ocean heat content above 3000 m for 1957–1994. While some studies have noted the potential importance of the choice of infilling method in poorly sampled regions (Gregory et al., 2004; AchutaRao et al., 2006), the consistency of the differently processed data from Levitus et al. (2005), Ishii et al. (2006) and Willis et al. (2004) analyses adds confidence to their use for analysing trends in climate change studies (Chapter 5). Gregory et al. (2004) show that agreement between models and observations is better in the well-observed upper ocean (above 300m) in the Northern Hemisphere and that there is large sensitivity to the method of infilling the observational dataset outside this well-observed region. They find a strong maximum in variability in the Levitus dataset at around 500 m depth that is not seen in HadCM3, a possible indication of model deficiency or alternatively an artefact in the Levitus data. AchutaRao et al. (2006) also find that the effects of sparse observational coverage and the method of infilling have significant impacts on the representativeness of the observationally-based estimates of variability over much of the oceans.

9.5.1.2 Water Mass Properties

Interior water masses, which are directly ventilated at the ocean surface, act to integrate highly variable surface changes in heat and freshwater, and could therefore provide indicators of global change (Stark et al., 2006). Some studies have attempted to investigate changes in 3-dimensional water mass properties (Chapter 5, Section 5.3). Sub-Antarctic mode water (SAMW) and the sub-tropical gyres have warmed in the Indian and Pacific basins since the 1960s, waters at high latitudes have freshened in the upper 500m and salinity has increased in some of the sub-tropical gyres (Chapter 5, Section 5.6). These changes are consistent with an increase in meridional moisture flux over the oceans over the last 50 years leading to increased precipitation at high latitudes (Chapter 5; Wong et al., 1999) and a reduction in the difference between precipitation and evaporation at mid-latitudes (Chapter 5, Section 5.6). This would suggest that the ocean might integrate rainfall changes to produce detectable salinity changes. Boyer et al. (2005) estimated linear trends of salinity for the global ocean from 1955 to 1998 indicating salinification in the Antarctic Polar Frontal Zone around 40°S and in the subtropical North Atlantic, and freshening in the sub-polar Atlantic (Chapter 5, Figures 5.5 and 5.7). However, variations in other terms (e.g. ocean fresh water transport) may be contributing substantially to the observed salinity changes and have not been quantified.

An observed freshening of Sub-Antarctic mode water (SAMW) in the South Indian Ocean between the 1960s and 1990s has been shown to be consistent with anthropogenically forced runs of HadCM3 (Banks et al., 2000) but care should be taken in interpreting sparse hydrographic data, since apparent trends could reflect natural variability or the aliased effect of changing observational coverage. Although SAMW was fresher on isopycnals in 1987 than in the 1960s, in 2002 the salinity was again near to the 1960s values (Bindoff and McDougall, 2000; Bryden et al., 2003). An analysis of an ocean model forced by observed atmospheric fluxes and sea surface temperatures indicates that this is likely associated with natural variability (Murray et al., 2006), a result supported by an analysis of 20th Century simulations with the HadCM3 model, which shows that it is not possible to reject the null hypothesis that the observed differences are due to internal variability (Stark et al., 2006), although the does project a long-term freshening trend in the 21st Century due to the large scale response to surface heating and hydrological changes (Banks et al., 2000).

9.5.1.3 Changes in the MOC

It is possible that anthropogenic and natural forcing may have influenced the meridional overturning circulation in the Atlantic (MOC; see also Chapter 5, Box 5.1). One possible oceanic consequence of climate change is a slowing down or even halting of the MOC. An estimate of the overturning circulation and associated heat transport based on a trans-Atlantic section along latitude 25°N indicates that the Atlantic MOC has slowed by about 30% over 5 samples taken between 1957 and 2004 (Bryden et al., 2005), although given the infrequent sampling and considerable variability it is not clear whether the trend estimate is robust (Chapter 5, Box 5.1). Freshening of North East Atlantic Deep Water has been observed (Dickson et al., 2002; Chapter 5, Figure 5.6; Curry et al., 2003) and has been interpreted as being consistent with an enhanced difference between precipitation and evaporation in high latitudes and a possible slowing down of the MOC. Wu et al. (2004) show that the observed freshening trend is well reproduced by an ensemble of HadCM3 simulations that includes both anthropogenic and natural forcings but this freshening coincides with a strengthening rather than a weakening trend in the MOC. Therefore this analysis is not consistent with an interpretation of the observed freshening trends in the North Atlantic as an early signal of a slow down of the thermohaline circulation. Dickson et al. (2002) propose a possible role for the Arctic in driving the observed freshening of the subpolar North Atlantic. Wu et al. (2005) show that observed increases in Arctic river flow (Peterson et al., 2002) are well simulated by HadCM3 including anthropogenic and natural forcings and propose that this increase is anthropogenic, since it is not seen in HadCM3 simulations including just natural forcing factors. However the relationship between this increased source of fresh water and freshening in the Labrador Sea is not clear in these HadCM3 simulations, since Wu et al. (2006) find that recent freshening in the Labrador Sea is seen in the model when it is driven by natural rather than anthropogenic forcings. Importantly, freshening is also associated with decadal and multi-decadal variability with links to the North Atlantic Oscillation (NAO; Box 5.1) and the Atlantic Multidecadal Oscillation (Chapter 5, Box 5.1; Vellinga and Wu, 2004; Knight et al., 2005).

9.5.2 Sea level

A precondition for attributing changes in sea level rise to anthropogenic forcing is that model-based estimates of historical global mean sea level rise should be consistent with observational estimates. Although AOGCM simulations of global mean surface air temperature trends are generally consistent with observations (Section 9.4.1, Figure 9.5), consistency for surface air temperature alone does not guarantee a realistic simulation of thermal expansion, as there may be compensating errors among climate sensitivity, ocean heat uptake and radiative forcing (cf. Raper et al., 2002, see also Section 9.6). Model simulations also offer the possibility of attributing past sea-level changes to particular forcing factors. The observational budget for sea level (Chapter 5, Section 5.5.6) assesses the periods 1961–2003 and 1993–2003. Table 9.2 evaluates the same terms from 20C3M simulations MMD (multi-model data archive at PCMDI); most 20C3M simulations end earlier (in 1999-2002), so the comparison is not quite exact.

[INSERT TABLE 9.2 HERE]

Simulations including natural as well as anthropogenic forcings (the "ALL" models in Table 9.2) generally have smaller ocean heat uptake during the period 1961-2003 than those without volcanic forcing since several large volcanoes cooled the climate during this period (Gleckler et al., 2006). This leads to a better agreement of those simulations with thermal expansion estimates based on observed ocean warming (Chapter 5, Section 5.5.3) than for the complete set of model simulations ("ALL/ANT" in Table 9.2). For 1993–2003 the models that include natural forcings agree well with observations. Although this result is somewhat uncertain because the simulations end at various dates from 1999 onwards, it accords with a result obtained by Church et al. (2005) using the PCM, and Gregory et al. (2006) using HadCM3, which suggest that 0.5 mm yr^{-1} of the trend in the last decade may result from warming as a recovery from the Pinatubo eruption of 1991. Comparison of the results for 1961–2003 and 1993–2003 shows that volcanoes influence the ocean differently over shorter and longer periods. The rapid expansion of 1993–2003 was caused, in part, by rapid warming of the upper ocean following the cooling due to the Pinatubo eruption, whereas the multi-decadal response is affected by the much longer persistence in the deep ocean of cool anomalies caused by volcanic eruptions (Delworth et al., 2005; Gleckler et al., 2006; Gregory et al., 2006).

Both observations and model results indicate that the global average mass balance of glaciers and ice caps depends linearly on global average temperature change, but observations of accelerated mass loss in recent years suggest a greater sensitivity than simulated by models. The global average temperature change simulated by AOGCMs gives a good match to the observational estimates of the contribution of glaciers and ice caps to sea level change in 1961–2003 and 1993–2003 (Table 9.2) with the assumptions that the global average mass balance sensitivity is 0.80 mm yr^{-1} K^{-1} (sea-level equivalent) and that the climate of 1900–1929 was 0.16 K warmer than the temperature required to maintain the steady state for glaciers (see discussion in Chapter 10, Section 10.6.3.1 and Appendix 10.A).

Calculations of ice-sheet surface mass balance changes due to climate change (following the methods of Gregory and Huybrechts (2006) and Section 10.6.3.1) indicate small but uncertain contributions during 1993–2003 of 0.1 ± 0.1 mm yr^{-1} (5-95% range) from Greenland and –0.2 ± 0.4 mm yr^{-1} from Antarctica, the latter being negative because rising temperature in AOGCMs leads to greater snow accumulation (but negligible melting) at present. The observational estimates (Chapter 4, Section 4.6.2 and Chapter 5, Section 5.5.6) are 0.21 ± 0.07 mm yr^{-1} for Greenland and 0.21 ± 0.35 mm yr^{-1} for Antarctica. For both ice sheets, there is a significant contribution from recent accelerations in ice flow leading to greater discharge of ice into the sea, an effect that is not included in the models because its causes and mechanisms are not yet properly understood (see Chapter 4, Section 4.6.2 and Chapter 10, Section10.6.4 for discussion). Hence the surface mass balance model underestimates the sea-level contribution from ice sheet melting. Model-based and observational estimates may also differ because the model-based estimates are obtained using estimates of the correlation between global mean climate change and local climate change over the ice sheets in the 21st century under SRES scenarios. This relationship may not represent recent changes over the ice sheets.

Summing the modelled thermal expansion, global glacier and ice cap, and the observational estimates of the ice sheet contributions results in totals that lie below the observed rates of global mean sea level rise during 1961–2003 and 1993–2003. As shown by Table 9.2, the terms are reasonably well reproduced by the models. Nevertheless, the discrepancy in the total, especially for 1961–2003, indicates the lack of a satisfactory explanation of sea-level rise. This is also a difficulty for the observational budget (discussed in Chapter 5, Section 5.5.6).

A discrepancy between model and observations could also be partly explained by the internally generated variability of the climate system, which control simulations suggest could give a standard deviation in the thermal expansion component of ~0.2 mm yr^{-1} in 10-year trends. This variability may be underestimated by models, since observations give a standard deviation in 10-year trends of 0.7 mm yr^{-1} in thermal expansion. (see Chapter 5, Section 5.5.3 and Section 9.5.1.1; Gregory et al., 2006).

Since recent warming and thermal expansion is likely largely anthropogenic (Section 9.5.1.1), the model results suggest that the greater rate of rise in 1993–2003 than in 1961–2003 could have been caused by rising anthropogenic forcing. However, tide-gauge estimates suggests larger variability than models in 10-year trends, and that rates as large as that observed during 1993–2003 occurred in previous decades (Chapter 5, Section 5.5.2.4).

Overall, it is very likely that the response to anthropogenic forcing contributed to sea level rise during the latter half of the 20th century. Models including anthropogenic and natural forcing simulate the observed thermal expansion since 1961 reasonably well. Anthropogenic forcing dominates the surface temperature change simulated by models, and has likely contributed to the observed warming of the upper ocean and widespread glacier retreat. It is very unlikely that the warming during the past half century is due only to known natural causes. Lack of studies quantifying the contribution of anthropogenic forcing to ocean heat content increase and glacier melting, and the fact that the observational budget is not closed, make it difficult to estimate the anthropogenic contribution. Nevertheless, an expert assessment based on modelling and ocean heat content studies suggests that anthropogenic forcing has likely contributed at least a quarter to half of the sea level rise during the second half of the 20th century (see also Woodworth et al., 2004).

Anthropogenic forcing is also expected to produce an accelerating rate of sea level rise (Woodworth et al., 2004). On the other hand, natural forcings could have increased the rate of sea-level rise in the early 20th century and decreased it later in the 20th century, thus producing a steadier rate of rise during the 20th century when combined with anthropogenic forcing (Crowley et al., 2003; Gregory et al., 2006). Observational evidence for acceleration during the 20th century is equivocal, but the rate of sea level rise was greater in the 20th than in the 19th century (Chapter 5, Section 5.5.2.4). An onset of higher rates of rise in the early 19th century could have been caused by natural factors, in particular the recovery from the Tambora eruption of 1815 (Crowley et al., 2003; Gregory et al., 2006), with anthropogenic forcing becoming important later in the 19th century.

9.5.3 Atmospheric Circulation Changes

Natural low frequency variability of the climate system is dominated by a small number of large scale circulation patterns such as the El Niño Southern Oscillation (ENSO), the Pacific Decadal Oscillation (PDO), and the Northern and Southern Annular Modes (NAM; SAM) (Chapter 3, Section 3.6 and Box 3.4). The impact of these modes on terrestrial climate on annual to decadal time scales can be profound, but the extent to which they can be excited or altered by external forcing remains uncertain. While some modes might be expected to change as a result of anthropogenic effects such as the enhanced greenhouse effect, there is little a priori expectation about the direction or magnitude of such changes.

9.5.3.1 El Niño Southern Oscillation/Pacific Decadal Oscillation

El Niño Southern Oscillation (ENSO) is the leading mode of variability in the tropical Pacific, and it has impacts on climate around the globe (Chapter 3, Section 3.6.2). There have been multidecadal oscillations in the ENSO index (conventionally defined as a mean SST anomaly in the eastern equatorial Pacific) throughout the 20th century, with more intense El Niño events since the late 1970s, which may reflect in part a mean warming of the eastern equatorial Pacific (Mendelssohn et al., 2005). Model projections of future climate change generally show a mean state shift towards more El-Niño-like conditions, with enhanced warming in the eastern tropical Pacific and a weakened Walker circulation (Chapter 10, Section 10.3.5.3); there is some evidence that such a weakening has been observed over the past 140 years (Vecchi et al., 2006). While some simulations of the response to anthropogenic influence have shown an increase in ENSO variability in response to greenhouse gas increases (Timmermann, 1999; Timmermann et al., 1999; Collins, 2000b), others have shown no change (e.g., Collins, 2000a), or a decrease in variability (Knutson et al., 1997). A recent survey of the simulated response to CO_2 doubling in fifteen MMD AOGCMs (Merryfield, 2006) found that three of the models exhibited significant increases in ENSO variability, five exhibited significant decreases and seven exhibited no significant change. Thus as yet there is no detectable change in ENSO variability in the observations, and no consistent picture of how it might be expected to change in response to anthropogenic forcing (Chapter 10, Section 10.3.5.4).

Decadal variability in the North Pacific is characterised by variations in the strength of the Aleutian Low coupled to changes in North Pacific SST (Chapter 3, Section 3.6.3 and Chapter 8, Section 8.4.2). The leading mode of decadal variability in the North Pacific is usually referred to as the PDO, and has a spatial structure in the atmosphere and upper North Pacific Ocean similar to the pattern that is associated with ENSO. One recent study showed a consistent tendency towards the positive phase of the PDO in observations and simulations with the MIROC model that included anthropogenic forcing (Shiogama et al., 2005), though differences between the observed and simulated PDO patterns, and the lack of additional studies, limit confidence in these findings.

9.5.3.2 North Atlantic Oscillation / Northern Annular Mode

The Northern Annular Mode (NAM) is an approximately zonally symmetric mode of variability in the Northern Hemisphere (Thompson and Wallace, 1998), and the North Atlantic Oscillation (NAO) (Hurrell, 1996) may be viewed as its Atlantic counterpart (Chapter 3, Section 3.6.4). The NAM index exhibited a pronounced trend towards its positive phase between the 1960s and the 1990s, corresponding to a decrease in surface pressure over the Arctic and an increase over the subtropical North Atlantic (see Chapter 3, Section 3.6.4; see also Hurrell, 1996; Thompson et al., 2000; Gillett et al., 2003a). Several studies have shown this trend to be inconsistent with simulated internal variability (Osborn et al., 1999; Gillett et al., 2000; Gillett et al., 2002b; Osborn, 2004; Gillett, 2005). Although the NAM index has decreased somewhat since its peak in the mid-1990s, the trend calculated over recent decades remains significant at the 5% level compared to simulated internal variability in most models (Osborn, 2004; Gillett, 2005), although one study found that the NAO index trend was marginally consistent with internal variability in one model (Selten et al., 2004).

Most climate models simulate some increase in the NAM index in response to increased concentrations of greenhouse gases (Fyfe et al., 1999; Paeth et al., 1999; Shindell et al., 1999; Gillett et al., 2003a; Gillett et al., 2003b; Osborn, 2004; Rauthe et al., 2004), although the simulated trend is generally smaller than that observed (Gillett et al., 2002b; Gillett et al., 2003b; Osborn, 2004; Gillett, 2005; and see Figure 9.16). Simulated sea level pressure changes are generally found to project more strongly onto the hemispheric NAM index than onto a two-station NAO index (Gillett et al., 2002b; Osborn, 2004; Rauthe et al., 2004). Some studies have postulated an influence of ozone depletion (Volodin and Galin, 1999; Shindell et al., 2001a), changes in solar irradiance (Shindell et al., 2001a), and volcanic eruptions (Kirchner et al., 1999; Shindell et al., 2001a; Stenchikov et al., 2006) on the NAM. Stenchikov et al. (2006) examined changes in sea level pressure following nine volcanic eruptions in the MMD 20C3M ensemble of 20th century simulations, and found that the models simulated a positive NAM response to the volcanoes, albeit one that was smaller than that observed. Nevertheless, ozone, solar and volcanic forcing changes are generally not found to have made a large contribution to the observed NAM trend over recent decades (Shindell et al., 2001a; Gillett et al., 2003a). Simulations incorporating all the major anthropogenic and natural forcings from the MMD 20C3M ensemble generally showed some increase in the NAM over the latter part of the 20th century (Gillett, 2005; Miller et al., 2006; and see Figure 9.16), though the simulated trend is in all cases smaller than that observed, indicating inconsistency between simulated and observed trends at the 5% significance level (Gillett, 2005).

The mechanisms underlying Northern Hemisphere circulation changes remain open to debate. Simulations in which observed SST changes, which may in part be externally forced, were prescribed either globally or in the tropics alone were able to capture around half of the recent trend towards the positive phase of the NAO (Hoerling et al., 2005; Hurrell et al., 2005), suggesting that the trend may in part relate to SST changes, particularly over the Indian Ocean (Hoerling et al., 2005). Another simulation in which a realistic trend in stratospheric winds was prescribed was able to reproduce the observed trend in the NAO (Scaife et al., 2005). Rind et al. (2005a; 2005b) find that both stratospheric changes and changes in SST can force changes in the NAM and NAO, with changes in SSTs being the dominant forcing mechanism.

Over the period 1968–1997, the trend in the NAM was associated with approximately 50% of the winter surface warming in Eurasia, due to increased advection of maritime air onto the continent, but only a small fraction (16%) of the NH extratropical annual mean warming trend (Thompson et al., 2000; Chapter 3, Section 3.6.4 and Figure 3.30). It was also associated with a decrease in winter precipitation over Southern

Europe and an increase over Northern Europe, due the northward displacement of the storm track (Thompson et al., 2000).

9.5.3.3 Southern Annular Mode

The Southern Annular Mode (SAM) is more zonally-symmetric than its Northern Hemisphere counterpart (Thompson and Wallace, 2000; Chapter 3, Section 3.6.5). It too has exhibited a pronounced upward trend over the past 30 years, corresponding to a decrease in surface pressure over the Antarctic and an increase over the Southern midlatitudes (Figure 9.16), although the mean SAM index since 2000 has been below the mean in the late 1990's, but above the long term mean (Figure 3.32). An upward trend in the SAM has occurred in all seasons, but the largest trend has been observed during the southern summer (Thompson et al., 2000; Marshall, 2003). Marshall et al. (2004) show that observed trends in the SAM are not consistent with simulated internal variability in HadCM3, suggesting an external cause. On the other hand, Jones and Widmann (2004) develop a 95-year reconstruction of the summer SAM index based largely on mid-latitude pressure measurements, and find that their reconstructed SAM index was as high in the early 1960's as in the late 1990's. However, a more reliable reconstruction from 1958 using more Antarctic data and a different method indicates that the summer SAM index was higher at the end of the 1990s than at any other time in the observed record (Marshall et al., 2004).

Based on an analysis of the structure and seasonality of the observed trends in Southern Hemisphere circulation, Thompson and Solomon (2002) suggest that they have been largely induced by stratospheric ozone depletion. Several modelling studies simulate an upward trend in the SAM in response to stratospheric ozone depletion (Sexton, 2001; Gillett and Thompson, 2003; Marshall et al., 2004; Shindell and Schmidt, 2004; Arblaster and Meehl, 2006; Miller et al., 2006), particularly in the southern summer. Stratospheric ozone depletion cools and strengthens the Antarctic stratospheric vortex in spring, and observations and models indicate that this strengthening of the stratospheric westerlies can be communicated downwards into the troposphere (Thompson and Solomon, 2002; Gillett and Thompson, 2003). While ozone depletion may be the dominant cause of the trends, other studies have indicated that greenhouse gas increases have also likely contributed (Fyfe et al., 1999; Kushner et al., 2001; Stone et al., 2001; Cai et al., 2003; Marshall et al., 2004; Shindell and Schmidt, 2004; Stone and Fyfe, 2005; Arblaster and Meehl, 2006) . During the Southern summer, the trend in the SAM has been associated with the observed increase in the circumpolar westerly winds over the Southern Ocean by ~3 ms^{-1}. This circulation change is estimated to explain most of the summer surface cooling over the Antarctic Plateau,, and about a third to a half of the warming of the Antarctic Peninsula (Thompson and Solomon, 2002; Carril et al., 2005; Chapter 3, Section 3.6.5), with the largest influence on the eastern side of the Peninsula (Marshall et al., 2006), though other factors are also likely to have contributed to this warming (Vaughan et al., 2001).

9.5.3.4 Sea Level Pressure Detection and Attribution

Global December–February sea level pressure changes observed over the past fifty years have been shown to be inconsistent with simulated internal variability (Gillett et al., 2003b; Gillett et al., 2005), but are consistent with the simulated response to greenhouse gas, stratospheric ozone, sulphate aerosol, volcanic aerosol and solar irradiance changes based on 20C3M simulations by eight MMD coupled models (Gillett et al., 2005) (Figure 9.16). This result is dominated by the Southern Hemisphere, where the inclusion of stratospheric ozone depletion leads to consistency between simulated and observed sea level pressure changes. In the Northern Hemisphere simulated sea level pressure trends are much smaller than those observed (Gillett, 2005). Global mean sea level pressure changes associated with increases in atmospheric water vapour are small in comparison to the spatial variations in the observed change in sea level pressure, and are hard to detect because of large observational uncertainties (Trenberth and Smith, 2005).

[INSERT FIGURE 9.16 HERE]

9.5.3.5 Monsoon Circulation

The current understanding of climate change in the monsoon regions remains one of considerable uncertainty with respect to circulation and precipitation (Chapter 3, Section 3.7; Chapter 8, Section 8.4.10; Chapter 10, Section 10.3.5.2). The Asian monsoon circulation in the MMD models was found to decrease by

15% by the late 21st century, under the SRES A1B scenario (Tanaka et al., 2005; Ueda et al., 2006), but trends during the 20th century were not examined. Ramanathan et al. (2005) simulated a pronounced weakening of the Asian monsoon circulation between 1985 and 2000 in response to black carbon aerosol increases. Chase et al. (2003) examined changes in several indices of four major tropical monsoonal circulations (Southeastern Asia, western Africa, eastern Africa, and the Australia/Maritime Continent) for the period 1950–1998. They found significantly diminished monsoonal circulation in each region, although this result is uncertain due to changes in the observing system affecting the NCEP reanalysis (Chapter 3, Section 3.7). These results are consistent with simulations (Ramanathan et al., 2005; Tanaka et al., 2005) of weakening monsoons due to anthropogenic factors, but further model and empirical studies are required to confirm this.

9.5.3.6 Tropical Cyclones

Several recent events, including the active North Atlantic hurricane seasons of 2004 and 2005, the unusual development of a cyclonic system in the subtropical South Atlantic that hit the coast of southern Brazil in March 2004 (e.g., Pezza and Simmonds, 2005), and a hurricane close to the Iberian Peninsula in October 2005, have raised public and media interest in the possible effects of climate change on tropical cyclone activity. The TAR concluded that there was "no compelling evidence to indicate that the characteristics of tropical and extratropical storms have changed", but that an increase in tropical peak wind intensities was likely to occur in some areas with an enhanced greenhouse effect (see also Chapter 3, Box 3.5 and Trenberth, 2005). The spatial resolution of most climate models limits their ability to realistically simulate tropical cyclones (Chapter 8, Section 8.5.3), and therefore most studies of projected changes in hurricanes have either used time slice experiments with high resolution atmosphere models and prescribed sea surface temperatures, or embedded hurricane models in lower resolution GCMs (Chapter 10, Section 10.3.6.3). While results vary somewhat, these studies generally indicate a reduced frequency of tropical cyclones in response to enhanced greenhouse gas forcing, but an increase in the intensity of the most intense cyclones (Chapter 10. Section 10.3.6.3). It has been suggested that the simulated frequency reduction may result from a decrease in radiative cooling associated with increased CO_2 concentration (Sugi and Yoshimura, 2004; Yoshimura and Sugi, 2005; Chapter 10, Section 10.3.6.3; Chapter 3, Box 3.5), while the enhanced atmospheric water vapour concentration under greenhouse warming increases available potential energy and thus cyclone intensity (Trenberth, 2005).

There continues to be little evidence of any trend in the observed total frequency of global tropical cyclones, at least up until the late 1990s (e.g., Solow and Moore, 2002; Elsener et al., 2004; Pielke et al., 2005; Webster et al., 2005). However, there is some evidence that tropical cyclone intensity may have increased. Globally, Webster et al. (2005) found a strong increase in the number and proportion of the most intense tropical cyclones over the past 35 years. Emanuel (2005) reported a marked increase since the mid-1970s of an index of the destructiveness of tropical cyclones (PDI; essentially an integral, over the lifetime of the cyclone, of the cube of the maximum wind speed) in the western North Pacific and North Atlantic, reflecting the apparent increases in both the duration of cyclones and their peak intensity. Several studies have shown that tropical cyclone activity was also high in the 1950–1970 period in the North Atlantic (Landsea, 2005) and North Pacific (Chan, 2006), although recent values of the PDI may be higher than those recorded previously (Emanuel, 2005; Chapter 3, Section 3.8.3). Emanuel (2005) and Elsner et al. (2006) report a strong correlation between the PDI and tropical Atlantic sea surface temperatures, although Chan and Liu (2004) find no analagous relationship in the western North Pacific. While changes in Atlantic sea surface temperatures have been linked in part to the Atlantic Multidecadal Oscillation, the recent warming appears to be mainly associated with increasing global temperatures (Chapter 3, Section 3.8.3.2; Mann and Emanuel, 2006; Trenberth and Shea, 2006). Tropical cyclone development is also strongly influenced by vertical wind shear and static stability (Chapter 3, Box 3.5). While increasing greenhouse gas concentrations have likely contributed to a warming of SSTs, effects on static stability and wind shear may have partly opposed this influence on tropical cyclone formation (Chapter 3, Box 3.5). Thus, detection and attribution of observed changes in hurricane intensity or frequency due to external influences remains difficult because of deficiencies in theoretical understanding of tropical cyclones, their modelling, and their long-term monitoring (e.g., Emanuel, 2005; Landsea, 2005; e.g., Pielke, 2005). These deficiencies preclude a stronger conclusion than an assessment that anthropogenic factors more likely than not have contributed to an increase in tropical cyclone intensity.

9.5.3.7 Extra-Tropical Cyclones

Simulations of 21st century climate change in the MMD 20C3M model ensemble generally exhibit a decrease in the total number of extra-tropical cyclones in both hemispheres, but an increase in the number of the most intense events (Lambert and Fyfe, 2006), although this behaviour is not reproduced in all models (Bengtsson et al., 2006; Chapter 10, Section 10.3.6.4). Many 21st century simulations also show a poleward shift in the storm tracks in both hemispheres (Bengtsson et al., 2006; Chapter 10, Section 10.3.6.4). Recent observational studies of winter Northern Hemisphere storms have found a poleward shift in storm tracks and increased storm intensity, but a decrease in total storm numbers, in the second half of the 20th century (Chapter 3, Section 3.5.3). Analysis of observed wind and significant wave height suggests an increase in storm activity in the Northern Hemisphere. In the Southern Hemisphere, the storm track has also shifted poleward, with increases in the radius and depth of storms, but decreases in their frequency. These features appear to be associated with the observed trends in the Southern and Northern Annular Modes. Thus simulated and observed changes in extra-tropical cyclones are broadly consistent, but an anthropogenic influence has not yet been detected, owing to large internal variability and problems due to changes in observing systems (Chapter 3, Section 3.5.3).

9.5.4 Precipitation

9.5.4.1 Changes in Atmospheric Water Vapour

The amount of moisture in the atmosphere is expected to increase in a warming climate (Trenberth et al., 2005) because saturation vapour pressure increases with temperature according to the Clausius-Clapeyron equation. Satellite (SSMI) measurements of water vapour since 1988 are of higher quality than either radiosonde or reanalysis data (Trenberth et al., 2005) and show a statistically significant upward trend in precipitable (column-integrated) water of 1.2 ± 0.3 % per decade averaged over the global oceans (Chapter 3, Section 3.4.2.1). Soden et al. (2005) demonstrated that the observed changes, including the upward trend, are well simulated in the GFDL atmospheric model when observed SSTs are prescribed (Figure 9.17). The simulation and observations show common low frequency variability, which is largely associated with ENSO. Soden et al. (2005) also demonstrated that upper-tropospheric changes in water vapour are realistically simulated by the model. Observed warming over the global oceans is likely largely anthropogenic (Figure 9.12), suggesting that anthropogenic influence has contributed to the observed increase in atmospheric water vapour over the oceans.

[INSERT FIGURE 9.17 HERE]

9.5.4.2 Global Precipitation Changes

The increased atmospheric moisture content associated with warming might be expected to lead to increased global mean precipitation (Section 9.5.4.1). Global annual land mean precipitation showed a small, but uncertain, upward trend over the 20th century of approximately 1.1 mm/decade (Chapter 3, Section 3.3.2.1 and Table 3.4). However, the record is characterised by large interdecadal variability, and global annual land mean precipitation shows a non-significant decrease since 1950 (Figure 9.18; see also Chapter 3, Table 3.4).

9.5.4.2.1 Detection of external influence on precipitation

Mitchell et al. (1987) argue that global mean precipitation changes should be controlled primarily by the energy budget of the troposphere where the latent heat of condensation is balanced by radiative cooling. Warming the troposphere enhances the cooling rate, thereby increasing precipitation, but this may be partly offset by a decrease in the efficiency of radiative cooling due to a CO_2 increase (Allen and Ingram, 2002; Yang et al., 2003; Lambert et al., 2004; Sugi and Yoshimura, 2004). This suggests that global mean precipitation should respond more to changes in shortwave forcing than CO_2 forcing, since shortwave forcings, such as volcanic aerosol, alter the temperature of the troposphere without affecting the efficiency of radiative cooling. This is consistent with a simulated decrease in precipitation following large volcanic eruptions (Robock and Liu, 1994; Broccoli et al., 2003), and may explain why anthropogenic influence has not been detected in measurements of global land mean precipitation (Ziegler et al., 2003; Gillett et al., 2004b), although Lambert et al. (2004) urge caution in applying the energy budget argument to land-only data. Greenhouse-gas induced increases in global precipitation may have also been offset by decreases due to anthropogenic aerosols (Ramanathan et al., 2001).

Several studies have demonstrated that simulated land mean precipitation in climate model integrations including both natural and anthropogenic forcings is significantly correlated with that observed (Allen and Ingram, 2002; Gillett et al., 2004b; Lambert et al., 2004), thereby detecting external influence in observations of precipitation (see Chapter 8, Section 8.3.1.2 for an evaluation of model simulated precipitation). Lambert et al. (2005) examine precipitation changes in simulations of nine MMD 20C3M models including anthropogenic and natural forcing (Figure 9.18a), and find that the responses to combined anthropogenic and natural forcing simulated by five of the nine models are detectable in observed land mean precipitation (Figure 9.18a). Lambert et al. (2004) detect the response to shortwave forcing, but not longwave forcing, in land mean precipitation using HadCM3, and Gillett et al. (2004b) similarly detect the response to volcanic forcing using the PCM. Climate models appear to underestimate the variance of land mean precipitation compared to that observed (Gillett et al., 2004b; Lambert et al., 2004; Lambert et al., 2005), but it is unclear whether this discrepancy results from an underestimated response to shortwave forcing (Gillett et al., 2004b), underestimated internal variability, errors in the observations, or a combination of these.

[INSERT FIGURE 9.18 HERE]

Greenhouse gas increases are also expected to cause enhanced horizontal transport of water vapour that is expected to lead to a drying of the subtropics and parts of the tropics (Kumar et al., 2004; Neelin et al., 2006), and a further increase in precipitation in the equatorial region and at high latitudes (Emori and Brown, 2005; Held and Soden, 2006). Simulations of twentieth century zonal mean land precipitation generally show an increase at high latitudes and near the equator, and a decrease in the subtropics of the Northern Hemisphere (Hulme et al., 1998; Figure 9.18b; Held and Soden, 2006). Projections of the 21st century show a similar effect (Chapter 10, Figure 10.12). This simulated drying of the northern subtropics and southward shift of the ITCZ may relate in part to the effects of sulphate aerosol (Rotstayn and Lohmann, 2002), although simulations without aerosol effects also show drying in the Northern subtropics (Hulme et al., 1998). This pattern of zonal mean precipitation changes is broadly consistent with that observed over the 20th century (Figure 9.18b; Hulme et al., 1998; Allen and Ingram, 2002; Rotstayn and Lohmann, 2002), although the observed record is characterized by large interdecadal variability (Chapter 3, Figure 3.15). The agreement between the simulated and observed zonal mean precipitation trends is not sensitive to the inclusion of forcing by volcanic eruptions in the simulations, suggesting that anthropogenic influence may be evident in this diagnostic.

Changes in runoff have been observed in many parts of the world, with increases or decreases corresponding to changes in precipitation (Chapter 3, Section 3.3.4). Climate models suggest that runoff will increase in regions where precipitation increases faster than evaporation, such as at high Northern latitudes (Chapter 10, Section 10.3.2.3 and Figure 10.12; see also Milly et al., 2005; Wu et al., 2005). Gedney et al. (2006) attributed increased continental runoff in the latter decades of the 20th century in part to suppression of transpiration due to CO_2-induced stomatal closure. They found that observed climate changes (including precipitation changes) alone were insufficient to explain the increased runoff, although their result is subject to considerable uncertainty in the runoff data. Also, Qian et al. (2006) simulate observed runoff changes in response to observed temperature and precipitation alone, and Milly et al. (2005) demonstrate that 20th century runoff trends simulated by the MMD models are significantly correlated with observed runoff trends. Wu et al. (2005) demonstrate that observed increases in Arctic river discharge are reproduced in coupled model simulations with anthropogenic forcing, but not in simulations with natural forcings only.

Mid-latitude summer drying is another anticipated response to greenhouse gas forcing (Chapter 10, Section 10.3.6.1), and drying trends have been observed in the both the Northern and Southern hemispheres since the 1950’s (Chapter 3, Section 3.3.4). Burke et al. (2006), using the HadCM3 model with all natural and anthropogenic external forcings and a global Palmer Drought Severity Index dataset compiled from observations by Dai et al. (2004), were able to formally detect the observed global trend towards increased drought in the second half of the 20th century, although the model trend was weaker than observed and the relative contributions of natural external forcings and anthropogenic forcings was not assessed. The model also simulated some aspects of the spatial pattern of observed drought trends, such as the trends across much of Africa and southern Asia, but not others, such as the trend to wetter conditions in Brazil and northwest Australia.

9.5.4.2.2 Changes in extreme precipitation

Allen and Ingram (2002) suggest that while global annual mean precipitation is constrained by the energy budget of the troposphere, extreme precipitation is constrained by the atmospheric moisture content, as predicted by the Clausius-Clapeyron equation. For a given change in temperature they therefore predict a larger change in extreme precipitation than in mean precipitation, which is consistent with the HadCM3 response. Consistent with these findings, Emori and Brown (2005) discuss physical mechanisms governing changes in the dynamic and thermodynamic components of mean and extreme precipitation and conclude that changes related to the dynamic component (i.e., that due to circulation change) are secondary factors in explaining the greater percentage increase in extreme precipitation than in mean precipitation that is seen in models. Meehl et al. (2005) demonstrate that tropical precipitation intensity increases are related to water vapour increases, while mid-latitude intensity increases are related to circulation changes that affect the distribution of increased water vapour.

Climatological data show that the most intense precipitation occurs in warm regions (Easterling et al., 2000) and diagnostic analyses have shown that even without any change in total precipitation, higher temperatures lead to a greater proportion of total precipitation in heavy and very heavy precipitation events (Karl and Trenberth, 2003). In addition, Groisman et al. (1999) have demonstrated empirically, and Katz (1999) theoretically, that as total precipitation increases a greater proportion falls in heavy and very heavy events if the frequency remains constant. Similar characteristics are anticipated under global warming (Cubasch et al., 2001; Semenov and Bengtsson, 2002; Trenberth et al., 2003). Trenberth et al. (2005) point out that since the amount of moisture in the atmosphere is likely to rise much faster as a consequence of rising temperatures than the total precipitation, this should lead to an increase in the intensity of storms, offset by decreases in duration or frequency of events.

Model results also suggest that future changes in precipitation extremes will likely be greater than changes in mean precipitation (Chapter 10, Section 10.3.6.1; see Chapter 8, Section 8.5.2 for an evaluation of model simulated precipitation extremes). Simulated changes in globally averaged annual mean and extreme precipitation appear to be quite consistent between models. The greater and spatially more uniform increases in heavy precipitation as compared to mean precipitation may allow extreme precipitation change to be more robustly detectable (Hegerl et al., 2004).

Evidence for changes in observations of short-duration precipitation extremes varies with the region considered (Alexander et al., 2006) and the analysis method that is employed (Folland et al., 2001; Chapter 3, Section 3.8.2.2). Significant increases in observed extreme precipitation have been reported over some parts of the world, for example over the United States, where the increase is similar to changes expected under greenhouse warming (e.g., Karl and Knight, 1998; Semenov and Bengtsson, 2002; Groisman et al., 2005). However, a quantitative comparison between area-based extreme events simulated in models and station data remains difficult because of the different scales involved (Osborn and Hulme, 1997). A first attempt based on Frich et al. (2002) indices used fingerprints from atmospheric model simulations with prescribed sea surface temperature (Kiktev et al., 2003). This study found little similarity between patterns of simulated and observed rainfall extremes. This is in contrast to the qualitative similarity found in other studies (Semenov and Bengtsson, 2002; Groisman et al., 2005). Tebaldi et al. (2006) reported that eight MMD 20C3M models showed a general tendancy towards a greater frequency of heavy-precipitation events over the past four decades, most coherently in the high latitudes of the Northern Northern Hemisphere, broadly consistent with observed changes (Groisman et al., 2005).

9.5.4.3 Regional Precipitation Changes

Observed trends in annual precipitation during the period 1901 to 2003 are shown in Chapter 3, Figure 3.13 for regions in which data is available. Responses to external forcing in regional precipitation trends are expected to exhibit low signal-to-noise ratios and are likely to exhibit strong spatial variations because of the dependence of precipitation on atmospheric circulation and on geographic factors such as orography. There have been some suggestions, for specific regions, of a possible anthropogenic influence on precipitation, which are discussed below.

9.5.4.3.1 Sahel drought

Rainfall decreased substantially across the Sahel from the 1950s until at least the late 1980s (Dai et al., 2004; Figure 9.19, see also Chapter 3, Figure 3.37). There has been a partial recovery since about 1990, although rainfall has not returned to levels typical of the period 1920–1965. Zeng (2003) noted that two main hypotheses have been proposed as a cause of the extended drought: overgrazing and conversion of woodland to agriculture increasing surface albedo and reducing moisture supply to the atmosphere, and large-scale atmospheric circulation changes related to decadal global sea surface temperature changes that could be of anthropogenic or natural origin (Nicholson, 2001). Black carbon has also been suggested as a contributor (Menon et al., 2002a). Taylor et al. (2002) examined the impact of land use change with an atmospheric GCM forced only by estimates of Sahelian land use change since 1961. They simulated a small decrease in Sahel rainfall (around 5% by 1996) and concluded that the impacts of recent land use changes are not large enough to have been the principal cause of the drought.

Several recent studies have demonstrated that simulations with a range of atmospheric models using prescribed observed SSTs are able to reproduce observed decadal variations in Sahel rainfall (Bader and Latif, 2003; Giannini et al., 2003; Rowell, 2003; Haarsma et al., 2005; Held et al., 2005; Lu and Delworth, 2005; see also Figure 9.19; Hoerling et al., 2006), consistent with earlier findings (Folland, 1986; Rowell, 1996). Hoerling et al. (2006) show that AGCMs with observed SST changes typically underestimate the magnitude of the observed precipitation changes, although the models and observations are not inconsistent. These studies differ somewhat in terms of which ocean SSTs they find to be most important: Giannini et al. (2003) and Bader and Latif (2003) emphasize the role of tropical Indian Ocean warming, Hoerling et al. (2006) attribute the drying trend to a progressive warming of the South Atlantic relative to the North Atlantic, and Rowell (2003) finds that Mediterranean SSTs are an additional important contributor to decadal variations of Sahel rainfall. Based on a multi-model ensemble of coupled model simulations Hoerling et al. (2006) concluded that the observed drying trend in the Sahel is not consistent with simulated internal variability alone.

[INSERT FIGURE 9.19 HERE]

Thus recent research indicates that changes in sea surface temperatures are probably the dominant influence on rainfall in the Sahel, although land use changes possibly also contribute (Taylor et al., 2002). But what has caused the differential SST changes? Rotstayn and Lohmann (2002) proposed that spatially-varying, anthropogenic sulphate aerosol forcing (both direct and indirect) can alter low-latitude atmospheric circulation leading to a decline in Sahel rainfall. They found a southward shift of tropical rainfall due to a hemispheric asymmetry in the sea surface temperature response to changes in cloud albedo and lifetime in a climate simulation forced with recent anthropogenic changes in sulphate aerosol. Williams et al. (2001) also found a southward shift of tropical rainfall as a response to the indirect effect of sulphate aerosol. These results suggest that sulphate aerosol changes may have led to reduced warming of the northern tropical oceans which in turn led to the decrease in Sahel rainfall, possibly enhanced through land-atmosphere interaction, though a full full attribution analysis has yet to be conducted. Held et al. (2005) showed that historical climate simulations with the both the GFDL-CM2.0 and CM2.1 models exhibit drying trends over the Sahel in the second half of the 20th century, which they ascribe to a combination of greenhouse gas and sulphate aerosol changes. The spatial pattern of the trends in simulated rainfall also shows some agreement with observations. However, Hoerling et al. (2006) found that eight other coupled climate models with prescribed anthropogenic forcing showed no significant trends in Sahel rainfall over the 1950-1999 period.

9.5.4.3.2 Southwest Australian drought

Early winter (May–July) rainfall in the far southwest of Australia declined by about 15% in the mid-1970s (IOCI, 2002) and remained low subsequently. The rainfall decrease was accompanied by a change in large-scale atmospheric circulation in the surrounding region (Timbal, 2004). The circulation and precipitation changes are somewhat consistent with, but larger than those simulated by climate models in response to greenhouse gas increases. IOCI (2005) concluded that land cover change could not be the primary cause of the rainfall decrease because of the link between the rainfall decline and changes in large-scale atmospheric circulation, and re-affirmed the conclusion of IOCI (2002) that both natural variability and greenhouse forcing likely contributed. Timbal et al. (2005) demonstrate that climate change signals downscaled from the PCM show some similarity to observed trends, although the significance of this finding is uncertain.

Some authors have suggested that the decrease in rainfall is related to anthropogenic changes in the Southern Annular Mode (SAM, see Section 9.5.3.3), (e.g., Karoly, 2003). However, the influence of change in circulation on southwest Australian drought remains unclear as the largest SAM trend has occurred during the southern hemisphere summer (December–March; Thompson et al., 2000; Marshall et al., 2004), while the largest rainfall decrease has occurred in early winter (May–July).

9.5.4.3.3 Monsoon precipitation

Decreasing trends in precipitation over the Indonesian Maritime Continent, equatorial western and central Africa, Central America, Southeast Asia, and eastern Australia have been observed over the period 1948–2003, while increasing trends were found over the United States and northwestern Australia (Section 3.7). The TAR (IPCC, 2001, pp 568) concluded that an increase in southeast Asian summer monsoon precipitation is simulated in response to greenhouse gas increases in climate models, but that this effect is reduced by an increase in sulphate aerosols, which tend to decrease monsoon precipitation. Since then, additional modelling studies have come to conflicting conclusions regarding changes in monsoon precipitation (Lal and Singh, 2001; Douville et al., 2002; Maynard et al., 2002; Wang and Lau, 2003; May, 2004; Wardle and Smith, 2004; see also Section 9.5.3.5). Ramanathan et al. (2005) were able to simulate realistic changes in Indian monsoon rainfall, particularly a decrease which occurred between 1950 and 1970, by including the effects of black carbon aerosol. In both the observations and model, these changes were associated with a decreased SST gradient over the Indian Ocean and an increase in tropospheric stability, and they were not reproduced in simulations with greenhouse gas and sulphate aerosol changes only.

9.5.5 Cryosphere Changes

9.5.5.1 Sea Ice

Widespread warming would, in the absence of other countervailing effects, lead to declines in sea ice, snow, and glacier and ice-sheet extent and thickness. The annual mean area of Arctic sea ice cover has decreased in recent decades, with stronger declines in summer than in winter, and some thinning (Chapter 4, Section 4.4). Gregory et al. (2002b) showed that a four-member ensemble of integrations of HadCM3 with all major anthropogenic and natural forcing factors, simulated a decline in Arctic sea ice extent of about 2.5% per decade over the period 1970–1999, which is close to the observed decline of 2.7% per decade over the satellite period 1978–2004. This decline is inconsistent with simulated internal climate variability and the response to natural forcings alone (Vinnikov et al., 1999; Gregory et al., 2002b; Johannssen et al., 2004), indicating that anthropogenic forcing has likely contributed to the trend in NH sea ice extent. Models such as those described by Rothrock et al. (2003) and references therein are able to reproduce the observed interannual variations in ice thickness, at least when averaged over fairly large regions. Simulations of historical Arctic ice thickness or volume (Goeberle and Gerdes, 2003; Rothrock et al., 2003) show a marked reduction in ice thickness starting in the late 1980s, but disagree somewhat with respect to trends and/or variations earlier in the century. Although some of the dramatic change inferred may be a consequence of a spatial redistribution of ice volume over time (e.g., Holloway and Sou, 2002), thermodynamic changes are also believed to be important. Low-frequency atmospheric variability (such as interannual changes in circulation connected to the Arctic Oscillation) appears to be important in flushing ice out of the Arctic Basin, thus increasing the amount of summer open water and enhancing thermodynamic thinning through the ice-albedo feedback (e.g., Lindsay and Zhang, 2005). Large-scale modes of variability affect both wind-driving and heat transport in the atmosphere, and therefore contribute to interannual variations in ice formation, growth and melt (e.g., Rigor et al., 2002; Dumas et al., 2003). Thus the decline in Arctic sea ice extent and its thinning appears to be largely, but not wholly, due to greenhouse gas forcing.

Unlike in the Arctic, a strong decline in sea ice extent has not been observed in the Antarctic during the period of satellite observations (Chapter 4, Section 4.4.2.2). Fichefet et al. (2003) conducted a simulation of Antarctic ice thickness using observationally-based atmospheric forcing covering the period 1958 to 1999. They noted pronounced decadal variability, with area-average ice thickness varying by ±0.1m (compared to a mean thickness of roughly 0.9m), but no long-term trend. However, Gregory et al. (2002b) found a decline in Antarctic sea ice extent in their model, contrary to observations. They suggested that the lack of consistency between the observed and modelled changes in sea ice extent might reflect an unrealistic simulation of regional warming around Antarctica, rather than a deficiency in the ice model. Holland and

Raphael (2006) examined sea-ice variability in six MMD 20C3M simulations that included stratospheric ozone depletion. Holland and Raphael (2006) conclude that the observed weak increase in Antarctic sea ice extent is not inconsistent with simulated internal variability, with some simulations reproducing the observed trend over 1979-2000, although the models exhibit larger interannual variability in sea ice extent than satellite observations.

9.5.5.2 Snow and Frozen Ground

Snow cover in the Northern Hemisphere, as measured from satellites, has declined substantially in the past 30 years, particularly from early spring through summer (Chapter 4, Section 4.2). Trends in snow depth and cover can be driven by precipitation or temperature trends. The trends in recent decades have generally been driven by warming at lower and middle elevations. Evidence for this includes: (a) Interannual variations in NH April snow-covered area are strongly correlated (r = –0.68) with April 40–60°N temperature; (b) Interannual variations in snow (water equivalent, depth or duration) are strongly correlated with temperature at lower- and middle-elevation sites in North America (Mote et al., 2005), Switzerland (Scherrer et al., 2004), and Australia (Nicholls, 2005); (c) Trends in snow water equivalent or snow depth show strong dependence on elevation or equivalently mean winter temperature, both in western North America and Switzerland (with stronger decreases at lower, warmer elevations where a warming is more likely to affect snowfall and snow melt); and (d) The trends in North America, Switzerland and Australia have been shown to be well explained by warming and cannot be explained by changes in precipitation. In some very cold places, increases in snow depth have been observed and have been linked to higher precipitation.

Widespread permafrost warming and degradation appears to be the result of increased summer air temperatures and changes in the depth and duration of snow cover (Chapter 4, Section 4.7.2). The thickness of seasonally-frozen ground has decreased in response to winter warming and increases in snow depth (Chapter 4, Section 4.7.3).

9.5.5.3 Glaciers, Ice Sheets and Ice Shelves

During the 20th century, glaciers have generally lost mass with the strongest retreats in the 1930s and 1940s and after 1990 (Chapter 4, Section 4.5). The widespread shrinkage appears to imply widespread warming as the probable cause (Oerlemans, 2005), although in the tropics changes in atmospheric moisture might be contributing (Chapter 4, Section 4.5.3). Over the last half century, both global mean winter accumulation and summer melting have increased steadily (Ohmura, 2004; Dyurgerov and Meier, 2005; Greene, 2005), and at least in the Northern Hemisphere, winter accumulation and summer melting correlate positively with hemispheric air temperature (Greene, 2005); the negative correlation of net balance with temperature indicates the primary role of temperature in forcing the respective glacier fluctuations.

There have been a few studies for glaciers in specific regions examining likely causes of trends. Mass balances for glaciers in western North America are strongly correlated with global mean winter (October-April) temperatures and the decline in glacier mass balance has paralleled the increase in temperature since 1968 (Meier et al., 2003). Reichert et al. (2002a) forced a glacier mass balance model for the Nigardsbreen and Rhône glaciers with downscaled data from an AOGCM control simulation and concluded that the rate of glacier advance during the 'Little Ice Age' could be explained by internal climate variability for both glaciers, but that the recent retreat cannot, implying that the recent retreat of both glaciers is probably due to externally forced climate change. As well, the thinning and accelerating of at some polar glaciers (e.g., Thomas et al., 2004) appears to be the result of ice sheet calving driven by oceanic and atmospheric warming (Chapter 4, Section 4.6.3.4).

Taken together, the ice sheets of Greenland and Antarctica are shrinking. Slight thickening in inland Greenland is more than compensated for by thinning near the coast (Chapter 4, Section 4.6.2.2). Warming is expected to increase low-altitude melting and high-altitude precipitation in Greenland: Altimetry data suggest that the former effect is dominant. However, because some portions of ice sheets respond only slowly to climate changes, past forcing may be influencing ongoing changes, complicating attribution of recent trends (Chapter 4, Section 4.6.3.2).

9.5.6 Summary

In the TAR, quantitative evidence for human influence on climate was based almost exclusively on atmospheric and surface temperature. Since then, anthropogenic influence has also been identified in a range of other climate variables, such as ocean heat content, atmospheric pressure, and sea ice extent, thereby contributing further evidence of an anthropogenic influence on climate, and improving confidence in climate models.

Observed changes in ocean heat content have now been shown to be inconsistent with simulated natural climate variability, but consistent with a combination of natural and anthropogenic influence both on a global scale, and in individual ocean basins. Models suggest a substantial anthropogenic contribution to sea level rise, but underestimate the actual rise observed. While some studies suggest that an anthropogenic increase in high latitude rainfall may have contributed to a freshening of the Arctic Ocean and North Atlantic deep water, these results are still uncertain.

There is no evidence that 20th century ENSO behaviour is distinguishable from natural variability. By contrast, there has been a detectable human influence on global sea level pressure. Both the Northern and Southern Annular Modes have shown significant trends. Models reproduce the sign but not magnitude of the Northern Annular Mode trend, and models including both greenhouse gas and ozone simulate a realistic trend in the Southern Annular Mode. Anthropogenic influence has not been detected on either tropical or extra-tropical cyclones, although the apparent increased frequency of intense tropical cyclones, and its relationship to ocean warming, is suggestive of an anthropogenic influence.

Simulations and observations of total atmospheric water vapour averaged over oceans agree closely when the simulations are constrained by observed sea surface temperatures, suggesting that anthropogenic influence has contributed to an increase in total atmospheric water vapour. However, global mean precipitation is controlled not by the availability of water vapour, but by a balance between the latent heat of condensation and radiative cooling in the troposphere. This may explain why human influence has not been detected in global precipitation, while the influence of volcanic aerosols has been detected. However, observed changes in the latitudinal distribution of land precipitation are suggestive of a possible human influence as is the observed increased incidence of drought as measured by the Palmer Drought Severity Index (PDSI). Observational evidence indicates that the frequency of the heaviest rainfall events has likely increased within many land regions in general agreement with model simulations that indicate that rainfall in the heaviest events is likely to increase in line with atmospheric water vapour concentration. Many atmospheric GCMs capture the observed decrease in Sahel rainfall when constrained by observed sea surface temperatures, although this decreasse is not simulated by most AOGCMs. One study found that an observed decrease in Asian monsoon rainfall could only be simulated in response to black carbon aerosol, although conclusions regarding the monsoon response to anthropogenic forcing differ.

Observed decreases in Arctic sea ice extent have been shown to be inconsistent with simulated internal variability, and consistent with the simulated response to human influence, but Southern Hemisphere sea ice extent has not declined. The decreasing trend in global snow cover and widespread melting of glaciers is consistent with a widespread warming. Anthropogenic forcing has likely contributed substantially to widespread glacier retreat during the 20th century.

9.6 Observational Constraints on Climate Sensitivity

This section assesses recent research that infers equilibrium climate sensitivity and transient climate response from observed changes in climate. *Equilibrium climate sensitivity* (ECS) is the equilibrium annual global mean temperature response to a doubling of equivalent CO_2 from preindustrial levels and is thus a measure of the strength of the climate system's eventual response to greenhouse gas forcing. *Transient climate response* (TCR) is the annual global mean temperature change at the time of CO_2 doubling in a climate simulation with a 1%/yr compounded increase in CO_2 concentration (see Glossary and Chapter 8, Section 8.6.2.1 for detailed definitions). TCR is a measure of the strength and rapidity of the climate response to greenhouse gas forcing, and depends in part on the rate at which the ocean takes up heat. While the direct temperature change that results from greenhouse gas can be calculated in a relatively straightforward manner, uncertain atmospheric feedbacks (Chapter 8, Section 8.6) lead to uncertainties in estimates of future climate change. The objective here is to assess estimates of equilibrium climate sensitivity and transient climate response that are based on observed climate changes, while Chapter 8 assesses of feedbacks individually. Inferences on climate sensitivity from observed climate *changes* complement approaches in which uncertain parameters in climate models are varied and assessed by evaluating the the resulting skill in reproducing observed *mean* climate (Chapter 10, Section 10.5.4.4). While observed climate changes have the advantage of being most clearly related to future climate change, the constraints they provide on climate sensitivity are not yet very strong, in part because of uncertainties in both climate forcing and the estimated response (Section 9.2). An overall summary assessment of equilibrium climate sensitivity and transient climate response, based on the ability of models to simulate climate change and mean climate, and on other approaches, is given in Chapter 10, Box 10.2. Note also that this section does not assess regional climate sensitivity or sensitivity to forcings other than CO_2.

9.6.1 Methods to Estimate Climate Sensitivity

The most straightforward approach to estimating climate sensitivity would be to relate an observed climate change to a known change in radiative forcing. Such an approach is strictly correct only for changes between equilibrium climate states. Climatic states that were reasonably close to equilibrium in the past are often associated with substantially different climates than the preindustrial or present climate, which is probably not in equilibrium (Hansen et al., 2005). An example is the climate of the Last Glacial Maximum (Chapter 6 and Section 9.3). However the climate's sensitivity to external forcing will depend on the mean climate state and the nature of the forcing, both of which affect feedback mechanisms (Chapter 8). Thus a directly derived estimate of the sensitivity derived from the ratio of response to forcing cannot be readily compared to the sensitivity of climate to a doubling of CO_2 under idealized conditions. An alternative approach, which has been pursued in most work reported here, is based on varying parameters in climate models that influence the ECS in those models, and then attaching probabilities to the different ECS values based on the realism of the corresponding climate change simulations. This ameliorates the problem of feedbacks being dependent on the climatic state, but depends on the assumption that feedbacks are realistically represented in models and that uncertainties in all parameters relevant for feedbacks are varied. Despite uncertainties, results from simulations of climates of the past and recent climate change (Sections 9.3–9.5) increase confidence in this assumption.

The ECS and TCR estimates discussed here are generally based on large ensembles of simulations using climate models of varying complexity, where uncertain parameters influencing the model's sensitivity to forcing are varied. Studies vary key climate and forcing parameters in those models, such as the equilibrium climate sensitivity, the rate of ocean heat uptake, and in some instances, the strength of aerosol forcing, within plausible ranges. ECS can be varied directly in simple climate models and in some earth system models of intermediate complexity (the so-called EMICS; see Chapter 8), and indirectly in more complex EMICs and AOGCMs by varying model parameters that influence the strength of atmospheric feedbacks, for example, in cloud parameterizations. Since studies estimating ECS and TCR from observed climate changes require very large ensembles of simulations of past climate change (ranging from several hundreds to thousands of members), they are often, but not always, performed with EMICs or EBMs.

The idea that underlies this approach is that the plausibility of a given combination of parameter settings can be determined from the agreement of the resulting simulation of historical climate with observations. This is typically evaluated by means of Bayesian methods (see Appendix 9.B for methods). Bayesian approaches constrain parameter values by combining prior distributions that account for uncertainty in the knowledge of parameter values with information about the parameters that is estimated from data (Kennedy and O'Hagan, 2001). The uniform distribution has been used widely as a prior, which enables comparison of constraints obtained from the data in different approaches. ECS ranges encompassed by the uniform prior must be limited due to computer time limiting the size of model ensembles, but generally cover the range considered possible by experts, such as from 0 to 10°C. Note that uniform priors on ECS, which only require an expert assessment of possible range, generally assign a higher prior belief to high sensitivity than, for example, non-uniform priors that depend more heavily on expert assessments (e.g., Forest et al., 2006). Also, Frame et al. (2005) point out that care must be taken when specifying the uniform prior distribution. For example, a uniform prior distribution on the climate feedback parameter (see Glossary) implies a non-uniform prior distribution on ECS due to the nonlinear relationship between the two parameters.

Since observational constraints on the upper bound of ECS are still weak (as will be shown below), these prior assumptions influence the resulting estimates. Frame et al. (2005) advocate sampling a flat prior in ECS if this is the target of the estimate, or in TCR if future temperature trends are to be constrained. In contrast, statistical research on the design and interpretation of computer experiments suggests the use of prior distributions on model input parameters (e.g., see Kennedy and O'Hagan, 2001; Goldstein and Rougier, 2004). In such Bayesian studies, it is generally good practice to explore the sensitivity of results to different prior beliefs (see, for example, Tol and Vos, 1998; O'Hagan and Forster, 2004). Furthermore, as demonstrated by Annan and Hargreaves (2005) and Hegerl et al. (2006a), multiple and independent lines of evidence about climate sensitivity from, for example, analysis of climate change at different times, can be combined by using information from one line of evidence as prior information for the analysis of another line of evidence. The extent to which the different lines of evidence provide complete information on the underlying physical mechanisms and feedbacks that determine the climate sensitivity is still an area of active research. In the following, uniform priors on the target of the estimate are used unless otherwise specified.

Methods that incorporate a more comprehensive treatment of uncertainty generally produce wider uncertainty ranges on the inferred climate parameters. Methods that do not vary uncertain parameters, such as ocean diffusivity, in the course of the uncertainty analysis will yield probability distributions for climate sensitivity that are conditional on these values, and therefore are likely to underestimate the uncertainty in climate sensitivity. On the other hand, approaches that do not use all available evidence will produce wider uncertainty ranges than estimates that are able to use observations more comprehensively.

9.6.2 Estimates of Climate Sensitivity Based on Instrumental Observations

9.6.2.1 Estimates of Climate Sensitivity Based on 20th Century Warming

A number of recent studies have used instrumental records of surface, ocean and atmospheric temperature changes to estimate climate sensitivity. Most studies use the observed surface temperature changes over the 20th century or the last 150 years (Chapter 3). In addition, some studies also use the estimated ocean heat uptake since 1955 based on Levitus et al. (2000; 2005) (Chapter 5), and temperature changes in the free atmosphere (Chapter 3; see also Table 9.3). For example, Frame et al. (2005) and Andronova and Schlesinger (2000) use surface air temperature alone, while Forest et al. (2002; 2006), Harvey and Kaufmann (2002), Knutti et al. (2002; 2003) and Gregory et al. (2002a) use both surface air temperature and ocean temperature change to constrain climate sensitivity. Forest et al. (2002; 2006) and Lindzen and Giannitsis (2002) use free atmospheric temperature data from radiosondes in addition to surface air temperature. Note that studies using radiosonde data may be affected by recently discovered inhomogeneities (Chapter 3, Section 3.4.1.1), although Forest et al. (2006) illustrate that the impact of the radiosonde atmospheric temperature data on their climate sensitivity estimate is smaller than that of surface and ocean warming data. A further recent study uses ERBE Earth radiation budget data (Forster and Gregory, 2006) in addition to surface temperature changes to estimate climate feedbacks (and thus ECS) from observed changes in forcing and climate.

[INSERT TABLE 9.3 HERE]

Wigley et al. (1997) pointed out that uncertainties in forcing and response made it impossible to use observed global temperature changes to constrain ECS more tightly than the range explored by climate models at the time (1.5 to 4.5°C), and particularly the upper end of the range, a conclusion confirmed by subsequent studies. A number of subsequent publications qualitatively describe parameter values that allow models to reproduce features of observed changes, but without directly estimating a climate sensitivity probability density function (pdf). For example, Harvey and Kaufmann (2002) find a best-fit ECS of 2.0°C out of a range of 1–5°C, and constrain fossil fuel and biomass aerosol forcing (Section 9.2.1.2). Lindzen and Giannitsis (2002) pose the hypothesis that the rapid change in tropospheric (850–300 hPa) temperatures around 1976 triggered a delayed response in surface temperature that is best modelled with a climate sensitivity of less than 1°C. However, their estimate does not account for substantial uncertainties in the analysis of such a short time period, most notably that which is associated with the role of internal climate variability in the rapid tropospheric warming of 1976. The 1976/1977 climate shift occurred along with a phase shift of the Pacific Decadal Oscillation, and a concurrent change in the ocean (see Chapter 3, Section 3.6.3) that appears to contradict the Lindzen and Giannitsis (2002) assumption that the change was initiated by tropospheric forcing. Also, the authors do not account for uncertainties in the simple model whose sensitivity is fitted. The finding of Lindzen and Giannitsis is in contrast with that of Forest et al. (2002; 2006) who considered the joint evolution of surface and upper air temperatures on much longer timescales.

Several recent studies have derived probability estimates for ECS using a range of models and diagnostics. The diagnostics, which are used to compare model simulated and observed changes, are often simple temperature indices such as the global mean surface temperature and ocean mean warming (Knutti et al., 2002; 2003) or the differential warming between the Southern and Northern Hemispheres (together with global mean, Andronova and Schlesinger, 2001). Results that use more detailed information about the space-time evolution of climate may be able to provide tighter constraints than those that use simpler indices. Forest et al. (2002; 2006) use a so called "optimal" detection method (Section 9.4.1.4 and Appendix 9.A.1) to diagnose the fit between model simulated and observed patterns of zonal mean temperature change. Frame et al. (2005) use detection results from an analysis based on several multi-model AOGCM fingerprints (Section 9.4.1.4) that separate the greenhouse gas response from that to other anthropogenic and natural forcings (Stott et al., 2006c). Similarly, Gregory et al. (2002a) apply an inverse estimate of the range of aerosol forcing based on fingerprint detection results. Note that while results from fingerprint detection approaches will be affected by uncertainty in separation between greenhouse gas and aerosol forcing, the resulting uncertainty in estimates of the near-surface temperature response to greenhouse gas forcing is relatively small (Sections 9.2.3 and 9.4.1.4).

A further consideration in assessing these results is the extent to which realistic forcing estimates were used, and whether forcing uncertainty was included. Most studies consider a range of anthropogenic forcing factors, including greenhouse gases and sulphate aerosol forcing, sometimes directly including the indirect forcing effect, such as Knutti et al., (2002; 2003), sometimes indirectly accounting for the indirect effect by using a wide range of direct forcing (for example, Andronova and Schlesinger, 2001; Forest et al., 2002; Forest et al., 2006). Many studies also consider tropospheric ozone (e.g., Andronova and Schlesinger, 2001; Knutti et al., 2002, 2003). Forest et al. (2006) demonstrate that the inclusion of natural forcing affects the estimated probability density function of climate sensitivity since net negative natural forcing in the second half of the 20th century favors higher sensitivities than earlier results disregarding natural forcing (Forest et al., 2002, see Figure 9.20), particularly if the same ocean warming estimates were used. Note that some of the changes due to inclusion of natural forcing were offset by using recently revised ocean warming data (Levitus et al., 2005), which favour somewhat smaller ocean heat uptakes than earlier data (Levitus et al., 2001; Forest et al., 2006). Only a few estimates account for uncertainty in forcings other than from aerosols (e.g. Gregory et al., 2002a; e.g. Knutti et al., 2002; Knutti et al., 2003); some other studies perform some sensitivity testing to assess the effect of forcing uncertainty not accounted for, for example, in natural forcing (e.g., Forest et al., 2006; see Table 9.1 for an overview).

The treatment of uncertainty in the ocean's uptake of heat varies, from assuming a fixed value for a model's ocean diffusivity (Andronova and Schlesinger, 2001) to trying to allow for a wide range of ocean mixing parameters (Knutti et al., 2002, 2003) or systematically varying the ocean's effective diffusivity (e.g., Forest et al., 2002; Frame et al., 2005; Forest et al., 2006). Furthermore, all approaches that use the climate's time evolution attempt to account for uncertainty due to internal climate variability, either by bootstrapping (Andronova and Schlesinger, 2001), by using a noise model in fingerprint studies whose results are used (Frame et al., 2005), or directly (Forest et al., 2002; Forest et al., 2006).

Figure 9.20 compares results from many of these studies. All pdfs shown are based on a uniform prior on ECS and have been rescaled to integrate to unity for all positive sensitivities up to 10°C to enable comparisons of results using different ranges of uniform priors (this affects both median and upper 95th percentiles if original estimates were based on a wider uniform range). Thus zero prior probability is assumed for sensitivities exceeding 10°C, since many results do not consider those, and also for negative sensitivities. Negative climate sensitivity would lead to cooling in response to a positive forcing and is inconsistent with with understanding of the energy balance of the system (Stouffer et al., 2000; Gregory et al., 2002a; Lindzen and Giannitsis, 2002). This figure shows that best estimates of the equilibrium climate sensitivity (mode of the estimated probability density functions) typically range between 1.2 and 4°C when inferred from constraints provided by historical instrumental data, in agreement with estimates derived from more comprehensive climate models. Most studies suggest a 5th percentile on climate sensitivity of 1°C or above. The upper 95th percentile is not well constrained, particularly in studies that account conservatively for uncertainty in, for example, 20th century radiative forcing and ocean heat uptake. The upper tail is particularly long in studies using diagnostics based on large-scale mean data because separation of the greenhouse gas response from that to aerosols or climate variability is more difficult with such diagnostics (Andronova and Schlesinger, 2001; Gregory et al., 2002a; Knutti et al., 2002, 2003). Forest et al. (2006) find a 5–95% range of 2.1 to 8.9°C for climate sensitivity (Table 9.3), which is a wider range than their earlier result based on anthropogenic forcing only (Forest et al., 2002). Frame et al. (2005) infer a 5–95% uncertainty range for the equilibrium climate sensitivity of 1.2 to 11.8°C, using a uniform prior distribution that extends well beyond 10°C sensitivity. Studies generally do not find meaningful constraints on the rate at which the climate system mixes heat into the deep ocean (e.g., Forest et al., 2002; Forest et al., 2006). However, Forest et al. (2002; 2006) find that many coupled AOGCMs mixed heat too rapidly into the deep ocean, which is broadly consistent with comparisons based on heat uptake (Section 9.5.1.1,). However the relevance of this finding is unclear because most MMD (multi-model data archive at PCMDI) AOGCMs were not included in the Forest et al. analysis, and because they used a relatively simple ocean model. Knutti et al. (2002) also determine that strongly negative aerosol forcing, as has been suggested by several observational studies (Anderson et al., 2003), is incompatible with the observed warming trend over the last century (Section 9.2.1.2 and Table 9.1).

Some studies have further attempted to use non-uniform prior distributions. Forest et al. (2002; 2006) obtained narrower uncertainty ranges when using expert prior distributions (Table 9.3). While they reflect credible prior ranges of ECS, expert priors may also be influenced by knowledge about observed climate change, and thus may yield overly confident estimates when combined with the same data (Appendix 9.B). Frame et al. (2005) find that sampling uniformly in TCR results in an estimated ECS of 1.2 to 5.2°C with a median value of 2.3°C. Also, several approaches have been based on a uniform prior distribution on climate feedback. Translating these results into ECS estimates is equivalent to using a prior distribution that favors smaller sensitivities, and hence tends to result in narrower ECS ranges (Frame et al., 2005). Forster and Gregory (2006) estimate ECS based on radiation budget data from the Earth Radiation Budget Experiment (ERBE) combined with surface temperature observations based on a regression approach, using the observation that there was little change in aerosol forcing over that time. They find a climate feedback parameter of 2.3 ± 1.4 W/(m^2 K), which corresponds to a 5–95% ECS range of 1.0 to 4.1°C if using a prior that puts more emphasis on lower sensitivities as discussed above, and a wider range if the prior is reformulated so that it is uniform in sensitivity (Table 9.3). The climate feedback parameter estimated from the MMD AOGCMs ranges from about 0.7 to 2.0 W/(m^2 K) (Supplementary material, Table S8.1).

[INSERT FIGURE 9.20 HERE]

9.6.2.2 Estimates Based on Individual Volcanic Eruptions

Some recent analyses have attempted to derive insights into ECS from the well observed forcing and response to the eruption of Mount Pinatubo, or from other major eruptions during the 20th century. Such events allow for the study of physical mechanisms and feedbacks and are discussed in detail in Section 8.6. For example, Soden et al. (2002) demonstrated agreement between observed and simulated responses based on an atmospheric GCM with a climate sensitivity of 3.0°C coupled to a mixed layer ocean, and that the agreement breaks down if the water vapour feedback in the model is switched off. Yokohata et al. (2005) find that a version of the MIROC climate model with a sensitivity of 4.0°C yields a much better simulation of the Mount Pinatubo eruption than a model version with sensitivity of 6.3°C, concluding that the cloud feedback in the latter model appears inconsistent with data. Note that both results may be specific to the model analyzed.

Constraining ECS from the observed responses to individual volcanic eruptions is difficult because the response to short-term volcanic forcing is strongly nonlinear in ECS, yielding only slightly enhanced peak responses and substantially extended response times for very high sensitivities (Frame et al., 2005; Wigley et al., 2005b). The latter are difficult to distinguish from a noisy background climate. A further difficulty arises from uncertainty in the rate of heat taken up by the ocean in response to a short, strong forcing. Wigley et al. (2005b) find that the lower boundary and best estimate obtained by comparing observed and simulated responses to major eruptions in the 20th century are consistent with the IPCC TAR range of 1.5 to 4.5°C, and that the response to the eruption of Mount Pinatubo suggests a best fit sensitivity of 3.0°C and an upper 95% limit of 5.2°C. However, as pointed out by the authors, this estimate does not account for forcing uncertainties. In contrast, an analysis by Douglass and Knox (2005) based on a box-model suggests a very low climate sensitivity (under 1°C) and negative climate feedbacks based on the eruption of Mount Pinatubo. Wigley et al. (2005a) demonstrate that the analysis method of Douglass and Knox (2005) severely underestimates climate sensitivity if applied to a model with known sensitivity (by a factor of 3). Furthermore, as pointed out by Frame et al. (2005), the effect of noise on the estimate of the climatic background level can lead to a substantial underestimate of uncertainties if not taken into account.

In summary, the responses to individual volcanic eruptions provide a useful test for feedbacks in climate models (Section 8.6). However, due to the physics involved in the response, such individual events cannot provide tight constraints on ECS. Estimates of the most likely sensitivity from most such studies are, however, consistent with those based on other analyses.

9.6.2.3 Constraints on Transient Climate Response

While ECS is the equilibrium global mean temperature change that eventually results from CO_2 doubling, the smaller *transient climate response* (TCR) refers to the global mean temperature change that is realized at the time of CO_2 doubling under an idealized scenario in which CO_2 concentrations increase at 1%/yr (Cubasch et al., 2001; see also Section 8.6.2.1). TCR is therefore indicative of the temperature trend associated with external forcing, and can be constrained by an observable quantity, the observed warming trend that is attributable to greenhouse gas forcing. Since external forcing is likely to continue to increase through the coming century, TCR may be more relevant to determining near term climate change than ECS.

Stott et al. (2006c) estimate TCR based on scaling factors for the response to greenhouse gases only (separated from aerosol and natural forcing in a 3-pattern optimal detection analysis) using fingerprints from three different model simulations (Figure 9.21) and find a relatively tight constraint. Using three model simulations together, their estimated median transient climate response is 2.1°C at the time of CO2 doubling (based on a 1% increase in CO2), with a 5–95% range of 1.5 to 2.8°C. Note that since TCR scales linearly with the errors in the estimated scaling factors, estimates do not show a tendency for a long upper tail, as is the case for ECS. However, the separation of greenhouse gas response from the responses to other external forcing in a multi-fingerprint analysis introduces a small uncertainty, illustrated by small differences in results between 3 models (Figure 9.21). TCR does not scale linearly with ECS because the transient response is strongly influenced by the speed with which the ocean transports heat into its interior, while the equilibrium sensitivity is governed by feedback strengths (discussion in Frame et al., 2005).

[INSERT FIGURE 9.21 HERE]

Estimates of a likely range for TCR can also be inferred directly from estimates of attributable greenhouse warming obtained in optimal detection analyses since there is a direct linear relationship between the two (Frame et al., 2005). The attributable greenhouse warming rates inferred from Figure 9.9 generally support the TCR range shown in Figure 9.21, although the lowest 5th percentile (1.3ºC) and the highest 95th percentile (3.3ºC) estimated in this way from detection and attribution analyses based on individual models lie outside the 5-95 percentile range of 1.5–2.8ºC obtained from Figure 9.21.

Choosing lower and upper limits that encompass the range of these results and deflating significance levels in order to account for structural uncertainty in the estimate leads to the conclusion that it is very unlikely that TCR is less than 1ºC and very unlikely that TCR is greater than 3.5ºC. Information based on the models discussed in Chapter 10 provides additional information that can help constrain TCR further (Chapter 10, Section 10.5.4.5).

9.6.3 Estimates of Climate Sensitivity Based on Paleoclimatic Data

The paleoclimate record offers a range of opportunities to assess the response of climate models to changes in external forcing. This section discusses estimates from both the paleoclimatic record of the last millennium, and from the climate of the Last Glacial Maximum (LGM). The latter gives a different perspective on feedbacks than anticipated with greenhouse warming, and thus provides a test bed for the physics in climate models. There also appears to be a likely positive relationship between temperature and CO_2, prior to the 650 000 year period covered by ice core measurements of CO_2 (Chapter 6, Section 6.3).

As with analyses of the instrumental record discussed in Section 9.6.2, some studies using paleoclimatic data have also estimated probability density functions for ECS by varying model parameters. Inferences about ECS made through direct comparisons between radiative forcing and climate response, without using climate models, show large uncertainties since climate feedbacks, and thus sensitivity, may be different for different climatic background states and for different seasonal characteristics of forcing (e.g., Montoya et al., 2000). Thus sensitivity to forcing during these periods cannot be directly compared to that for CO_2 doubling.

9.6.3.1 Estimates of Climate Sensitivity Based on Data for the Last Millennium

The relationship between forcing and response based on a long time horizon can be studied using paleo-climatic reconstructions of temperature and radiative forcing, particularly volcanism and solar forcing, for the last millennium. However, both forcing and temperature reconstructions are subject to large uncertainties (Chapter 6). To account for the uncertainty in reconstructions, Hegerl et al. (2006a) use several proxy data reconstructions of Northern Hemispheric extratropical temperature for the past millennium (Briffa et al., 2001; Esper et al., 2002; Mann and Jones, 2003; Hegerl et al., 2006b) to constrain ECS estimates for the pre-industrial period up to 1850. This study used a large ensemble of simulations of the last millennium performed with an energy balance model forced with reconstructions of volcanic (Crowley, 2000, updated), solar (Lean et al., 2002), and greenhouse gas forcing (Section 9.3.3 for results on the detection of these external influences). Their estimated probability density functions for ECS incorporate an estimate of uncertainty in the overall amplitude (including an attempt to account for uncertainty in efficacy), but not the time evolution, of volcanic and solar forcing. They also attempt to account for uncertainty in the amplitude of reconstructed temperatures in one reconstruction (Hegerl et al., 2006b), and assess the sensitivity of their results to changes in amplitude for others. All reconstructions combined yield a median climate sensitivity of 3.4°C and a 5–95% range of 1.2 to 8.6°C (Figure 9.20). Reconstructions with a higher amplitude of past climate variations (such as, Esper et al., 2002 or Hegerl et al., 2006b) are found to support higher ECS estimates than reconstructions with lower amplitude (e.g., Mann and Jones, 2003). Note that the constraint on ECS originates mainly from low-frequency temperature variations associated with changes in the frequency and intensity of volcanism which lead to a highly significant detection of volcanic response (Section 9.3.3) in all records used in the study.

The results of Andronova et al. (2004) are broadly consistent with these estimates. Andronova et al. (2004) demonstrate that climate sensitivities in the range of 2.3 to 3.4°C yield reasonable simulations of both the Northern Hemispheric mean temperature from 1500 onward when compared to the (Mann and Jones, 2003) reconstruction, and for the instrumental period. The agreement is less good for reconstructed Southern Hemisphere temperature, where reconstructions are substantially more uncertain (Chapter 6).

Rind et al. (2004) studied the period from about 1675 to 1715 to attempt a direct estimate of climate sensitivity. This period has reduced radiative forcing relative to the present due to decreased solar radiation, decreased greenhouse gas and possibly increased volcanic forcing (Section 9.2.1.3). Different Northern Hemisphere temperature reconstructions (Chapter 6, Figure 6.10) have a wide range of cooling estimates relative to the late 20th century that is broadly reproduced by climate model simulations. While climate in this cold period may have been close to radiative balance (Rind et al., 2004), some of the forcing during the present period is not yet realized in the system (estimated as 0.85 W m^{-2}, Hansen et al., 2005). Thus ECS estimates based on a comparison between radiative forcing and climate response are subject to large uncertainties, but are broadly similar to estimates discussed above. Again, reconstructions with stronger cooling in this period imply higher climate sensitivities than those with weaker cooling (results updated from Rind et al., 2004).

9.6.3.2 Inferences About Climate Sensitivity Based on the Last Glacial Maximum

The Last Glacial Maximum (LGM) is one of the key periods used to estimate equilibrium climate sensitivity (Hansen et al., 1984; Lorius et al., 1990; Hoffert and Covey, 1992), since it represents a quasi-equilibrium climate response to substantially altered boundary conditions. AOGCMs or EMICs identical or similar to those used for 20th and 21st century simulations, when forced with changes in greenhouse gas concentrations and the extent and height of ice sheet boundary conditions produce a 3.5–5.2°C cooling for this period in response to radiative perturbations of 4.1–7.2 W m^{-2} (Chapter 6, Section 6.4.1 and Section

9.3.3, see also Masson-Delmotte et al., 2006). The simulated cooling in the tropics ranges from 1.1 to 2.3°C. The equilibrium climate sensitivity (ECS) of the models used in the second phase of the Paleoclimate Model Intercomparison (PMIP2) ranges from 2.5 to 4°C (Chapter 8, Table 8.2), and there is some tendency that models with larger sensitivity produce larger tropical cooling for the LGM, but this relationship is not very tight. Comparison between simulated climate change and reconstructed climate is affected by substantial uncertainties in forcing and data (Chapter 6 and Section 9.2.1.3). For example, the PMIP2 forcing does not account for changes in mineral dust, since the level of scientific understanding for this forcing is very low (Chapter 6, Figure 6.5). The range of simulated temperature changes is also affected by differences of the radiative influence of the ice-covered regions in different models (Taylor et al., 2000). Nevertheless, the PMIP2 models simulate LGM climate changes that are approximately consistent with proxy information Chapter 6).

Recent studies (Annan et al., 2005; Schneider von Deimling et al., 2006) have attempted to estimate the probability density function of ECS from ensemble simulations of the LGM by systematically exploring model uncertainty. Both studies investigate the relationship between climate sensitivity and LGM tropical sea surface temperatures, which are influenced strongly by CO_2 changes. In a perturbed perturbed physics ensemble, Schneider von Deimling et al. (2006) vary 11 ocean and atmospheric parameters in a 1,000 member ensemble simulation of the LGM with the CLIMBER-2 EMIC (Chapter 8, Table 8.3). They found a close relationship between ECS and tropical SST cooling in their model, implying a 5–95% range of ECS of 1.2 to 4.3°C when attempting to account for model parameter, forcing and paleo-climate data uncertainties. Similar constraints on climate sensitivity were found when proxy reconstructions of LGM Antarctic temperatures were used instead of tropical SSTs (Schneider von Deimling et al., 2006). In contrast, Annan et al. (2005) used a perturbed physics ensemble based on a low-resolution version of the atmospheric component of the MIROC3.2 model, perturbing a range of model parameters over prior distributions determined from the ability of the model to reproduce seasonal mean climate in a range of climate variables. They find a best-fit sensitivity of about 4.5°C, and their results suggest that sensitivities in excess of 6°C are unlikely given observational estimates of LGM tropical cooling and the relationship between tropical SST and sensitivity in their model. Since the perturbed physics ensemble based on that atmospheric model does not produce sensitivities under 4°C, this result cannot provide a lower limit or a pdf for ECS.

The discrepancy between the inferred upper limits in the two studies probably arises from both different radiative forcing and structural differences between the models used. Forcing from changes in vegetation cover and dust is not included in the simulations done by (Annan et al., 2005), which would according to Schneider von Deimling et al., (2006) reduce the Annan et al. ECS estimates and yield better agreement between the results of the two studies. However, the effect of these forcings and their interaction with other LGM forcings is very uncertain, limiting confidence in such estimates of their effect (Chapter 6, Figure 6.5). Structural differences in models are likely to also play a role. The Annan et al. (2005) estimate shows a weaker association between simulated tropical SST changes and ECS than the Schneider von Deimling et al. (2006) result. Since Annan et al. use a mixed layer ocean model, and Schneider von Deimling a simplified ocean model, both models may not capture the full ocean response affecting tropical SSTs. The atmospheric model used in Schneider von Deimling is substantially simpler than that used in the Annan et al. (2005) study. Overall, estimates of climate sensitivity from the LGM are broadly consistent with other estimates of climate sensitivity derived, for example, from the instrumental period.

9.6.4 Summary of Observational Constraints for Climate Sensitivity

Any constraint of climate sensitivity obtained from observations must be interpreted in light of the underlying assumptions. These assumptions include (i) the choice of prior distribution for each of the model parameters (Section 9.6.1 and Appendix 9.B), including the parameter range explored, (ii) the treatment of other parameters that influence the estimate, such as effective ocean diffusivity, and (iii) the methods used to account for uncertainties, such as structural and forcing uncertainties, that are not represented by the prior distributions. Neglecting important sources of uncertainty in these estimates will result in overly narrow ranges that overstate the certainty with which the equilibrium climate sensitivity or transient climate response is known. Errors in assumptions about forcing or model response will also result in unrealistic features of model simulations, which can result in erroneous modes (peak probabilities) and shapes of the probability density function. On the other hand, using less than all available information will yield results that are less constrained than they could be under optimal use of available data.

While a variety of important uncertainties (e.g., radiative forcing, mixing of heat into the ocean) have been taken into account in most studies (Table 9.3), some caveats remain. Some processes and feedbacks might be poorly represented or missing, particularly, but not only, in simple or many intermediate complexity models. Structural uncertainties in the models, for example, in the representation of cloud feedback processes (Chapter 8) or the physics of ocean mixing, will affect results on climate sensitivity and are very difficult to quantify. Also, differences in efficacy between forcings are not directly represented in simple models, so they may affect the estimate (e.g., Tett et al., 2006), although this uncertainty may be folded into forcing uncertainty (e.g., Hegerl et al., 2003; 2006b). The use of a single value for the equilibrium climate sensitivity (ECS) further assumes that it is constant in time. However some authors (e.g., Senior and Mitchell, 2000; Boer and Yu, 2003) have shown that ECS varies in time in the climates simulated by their models. Since results from instrumental data and the last millennium are dominated primarily by decadal to century scale changes, they will therefore only represent climate sensitivity at an equilibrium that is not too far from the present climate. There is also a small uncertainty in the radiative forcing due to CO_2 doubling (<10%; see Chapter 2), which is not accounted for in most studies which derive observational constraints on climate sensitivity.

Despite these uncertainties, which are accounted for to differing degrees in the various studies, confidence is increased by the similarities between individual ECS estimates (Figure 9.20). Most studies find a lower 5% limit of between 1°C and 2.2°C, and studies that use information in a relatively complete manner generally find a most likely value between 2 and 3°C (Figure 9.20). Constraints on the upper end of the likely range of climate sensitivities are also important, particularly for probabilistic forecasts of future climate with constant radiative forcing. The upper 95% limit for ECS ranges from 5°C to 10°C, or greater in different studies depending upon the approach taken, the number of uncertainties included, and specific details of the prior distribution that was used. This wide range is largely caused by uncertainties and nonlinearities in forcings and response. For example, a high sensitivity is difficult to rule out because a high aerosol forcing could nearly cancel greenhouse gas forcing over the 20th century. This problem can be addressed, at least to some extent, if the differences in the spatial and temporal patterns of response between aerosol and greenhouse gas forcing are used for separating these two responses in observations (as, for example, in Gregory et al., 2002a; Harvey and Kaufmann, 2002; Frame et al., 2005). Also, nonlinearities in the response to transient forcing make it more difficult to constrain the upper limit on ECS based on observed transient forcing responses (Frame et al., 2005). The transient climate response, which may be more relevant for near-term climate change, is easier to constrain since it relates more linearly to observables. For the pre-instrumental part of the last millennium, uncertainties in temperature and forcing reconstructions, and the nonlinear connection between ECS and the response to volcanism prohibit tighter constraints. Estimates of climate sensitivity based on the ability of climate models to reproduce climatic conditions of the Last Glacial Maximum climate broadly support the ranges found from the instrumental period, although a tight constraint is also difficult to obtain from this period alone because of uncertainties in tropical temperature changes, forcing uncertainties, and the effect of structural model uncertainties. Also, the number of studies providing estimates of probability density functions from paleoclimatic data, using independent approaches and complementary sources of proxy data, are limited.

Thus most studies that use a simple uniform prior on equilibrium climate sensitivity are not able to exclude values beyond the traditional IPCC FAR range of 1.5 to 4.5°C (IPCC, 1990). However, considering all available evidence on ECS together provides a stronger constraint than individual lines of evidence. Bayesian methods can be used to incorporate multiple lines of evidence to sharpen the posterior distribution on ECS, as in Annan and Hargreaves (2006) and Hegerl et al. (2006a). Annan and Hargreaves (2006) demonstrate that using three lines of evidence, namely 20th century warming, the response to individual volcanic eruptions, and the LGM response, results in a tighter estimate of ECS, with a probability of less than 5% that equilibrium climate sensitivity exceeds 4.5°C. The authors find a similar constraint if using five lines of evidence under more conservative assumptions about uncertainties (adding cooling during the Little Ice Age and studies based on varying model parameters to match climatological means, see Box 10.2). However, as discussed in Annan and Hargreaves (2006), combining multiple lines of evidence may produce overly confident estimates unless every single line of evidence is entirely independent of others, or dependence is explicitly taken into accounted. Hegerl et al. (2006a) argue that instrumental temperature change during the second half of the 20th century is essentially independent of the paleo record of the last millennium and of the instrumental data from the first half of the 20th century that is used to calibrate the paleo records. Hegerl et al. (2006a) therefore base their prior probability distribution for the climate sensitivity on results from the late 20th century (Frame et al., 2005), which reduces the 5–95% ECS range from all proxy reconstructions analysed to 1.5 to 6.2°C compared to the previous range of 1.2 to 8.6°C. Both results demonstrate that independent estimates, when properly combined in a Bayesian analysis, can provide a tighter constraint on climate sensitivity, even if they individually provide only weak constraints. These studies also find a 5% lower limit of 1.5°C or above, consistent with several studies based on the 20th century climate change alone (Knutti et al., 2002; Forest et al., 2006) and estimates that greenhouse warming contributes substantially to observed temperature changes (Section 9.4.1.4).

Overall, several lines of evidence strengthen confidence in present estimates of equilibrium climate sensitivity, and new results based on objective analyses make it possible to assign probabilities to ranges of climate sensitivity previously assessed from expert opinion alone. This presents a significant advance. Results from studies of observed climate change and the consistency of estimates from different time periods indicate that equilibrium climate sensitivity is very likely larger than 1.5°C with a most likely value between 2 and 3°C. The lower bound is consistent with the view that the sum of all atmospheric feedbacks affecting climate sensitivity is positive. Although upper limits can be obtained by combining multiple lines of evidence, remaining uncertainties that are not accounted for in individual estimates (such as structural model uncertainties) and possible dependencies between individual lines of evidence make the upper 95% limit of equilibrium climate sensitivity uncertain at present. Nevertheless, constraints from observed climate change support the overall assessment that the equilibrium climate sensitivity is likely to lie between 2 to 4.5°C with a most likely value of approximately 3°C (Chapter 10, Box 10.2).

9.7 Combining Evidence of Anthropogenic Climate Change

The widespread change detected in temperature observations of the surface (Sections 9.4.1, 9.4.2, 9.4.3), free atmosphere (Section 9.4.4), and ocean (Section 9.5.1), together with consistent evidence of change in other parts of the climate system (Section 9.5), strengthens the conclusion that greenhouse gas forcing is the dominant cause of warming during the past several decades. This combined evidence, which is summarized in Table 9.4, is substantially stronger than the evidence that is available from observed changes in global surface temperature alone (Chapter 3, Figure 3.6).

[INSERT TABLE 9.4 HERE]

The evidence from surface temperature observations is strong: The observed warming is highly significant relative to estimates of internal climate variability which, while obtained from models, are consistent with estimates obtained from both instrumental data and paleo climate reconstructions. It is extremely unlikely (<5%) that recent global warming is due to internal variability alone such as might arise from El Niño (Section 9.4.1). The widespread nature of the warming (Chapter 3, Figures 3.9 and Figure 9.6) reduces the possibility that the warming could have resulted from internal variability. No known mode of internal variability leads to such widespread, near universal warming as has been observed in the past few decades. Although modes of internal variability such as El Niño can lead to global average warming for limited periods of time, such warming is regionally variable, with some areas of cooling (Chapter 3, Figures 3.27 and 3.28). Also, paleo-climatic evidence indicates that El Niño variability during the 20th century is not unusual relative to earlier periods (Section 9.3.3.2; Chapter 6). Paleo-climatic evidence suggests that such a widespread warming has not been observed in the Northern Hemisphere in at least the past 1200 years (Osborn and Briffa, 2006) further strengthening the evidence that the recent warming is not due to natural internal variability. Moreover, the response to anthropogenic forcing is detectable on all continents individually except Antarctica, and in some sub-continental regions. Climate models only reproduce the observed 20th century global mean surface warming when both anthropogenic and natural forcings are included (Figure 9.5). No model that has used natural forcing only has reproduced the observed global mean warming trend or the continental mean warming trends in all individual continents (except Antarctica) over the second half of the 20th century. Detection and attribution of external influences on 20th century and paleoclimatic reconstructions, from both natural and anthropogenic sources (Figure 9.4, Table 9.4), further strengthens the conclusion that the observed changes are very unusual relative to internal climate variability.

The energy content change associated with the observed widespread warming of the atmosphere is small relative to the energy content change of the ocean, and also smaller than that associated with other components such as the cryosphere. In addition, the solid earth also shows evidence for warming in boreholes (Huang et al., 2000; Beltrami et al., 2002; Pollack and Smerdon, 2004). It is theoretically feasible that the warming of the near surface could have occurred due to a reduction in the heat content of another component of the system. However, all parts of the cryosphere (glaciers, small ice caps, ice sheets, and sea-ice) have decreased in extent over the past half-century, consistent with anthropogenic forcing (Section 9.5.4, Table 9.4), implying that the cryosphere consumed heat and thus indicating that it could not have provided heat for atmospheric warming. More importantly, the heat content of the ocean (the largest reservoir of heat in the climate system) also increased, much more substantially than that of the other components of the climate system (Chapter 5, Figure 5.4; Hansen et al., 2005; Levitus et al., 2005). The warming of the upper ocean during the latter half of the 20th century was likely due to anthropogenic forcing (Barnett et al., 2005, Section 9.5.1, Table 9.4). While the statistical evidence in this research is very strong that the warming cannot be explained by ocean internal variability as estimated by two different climate models, uncertainty arises since there are discrepancies between estimates of ocean heat content variability from models and observations, although poor sampling of parts of the world ocean may explain this discrepancy. However, the spatial pattern of ocean warming with depth is very consistent with heating of the ocean resulting from net positive radiative forcing, since the warming proceeds downwards from the upper layers of the ocean; and there is deeper penetration of heat at middle to high latitudes and shallower penetration at low latitudes (Barnett et al., 2005; Hansen et al., 2005). This observed ocean warming pattern is inconsistent with a redistribution of heat between the surface and the deep ocean.

Thus the evidence appears to be inconsistent with the ocean or land being the source of the warming at the surface. Also, simulations forced with observed SST changes cannot fully explain the warming in the troposphere without increases in greenhouse gases (e.g., Sexton et al., 2001), further strengthening the evidence that the warming does not originate from the ocean. Further evidence for forced changes arises from widespread melting of the cryosphere (Section 9.5.5), increase in water vapour in the atmosphere (Section 9.5.4.1), and changes in top-of-the atmosphere radiation that are consistent with changes in forcing.

The simultaneous increase in energy content of all the major components of the climate system, and the pattern and amplitude of warming in the different components, together with evidence that the second half of the 20th century was likely the warmest in a 1000 years (Chapter 6) indicates that the cause of the warming is extremely unlikely to be the result of internal processes alone. The consistency across different lines of evidence makes a strong case for a significant human influence on observed warming at the surface. The observed rates of surface temperature and ocean heat content change are consistent with the understanding of the likely range of climate sensitivity and net climate forcings. Only by a net positive forcing, consistent with observational and model estimates of the likely net forcing of the climate system (as used in Figure 9.5), is it possible to explain the large increase in heat content of the climate system that has been observed (Chapter 5, Figure 5.4).

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Frequently Asked Question 9.1: Can Individual Extreme Events be Explained by Greenhouse Warming?

Changes in climate extremes are expected as the climate warms in response to increasing greenhouse gases from human activities, such as the use of fossil fuels. However, determining whether a specific, single extreme event is due to a specific cause such as increasing greenhouse gases, is difficult, if not impossible, for two reasons: 1) extreme events are usually caused by a combination of factors, and 2) a wide range of extreme events is a normal occurrence even in an unchanging climate. Nevertheless, analysis of the warming observed over the past century suggests that the likelihood of some extreme events, such as heat waves, has increased due to greenhouse warming, and that the likelihood of others, such as frost or extremely cold nights, has decreased. For example, a recent study estimates that human influences have more than doubled the risk of a very hot European summer like that of 2003.

People affected by an extreme weather event often ask whether human influences on the climate could be held to some extent responsible. Recent years have seen many extreme events that some commentators have linked to increasing greenhouse gases. These include the prolonged drought in Australia, the extremely hot summer in Europe in 2003 (see Figure 1), the intense North Atlantic hurricane seasons of 2004 and 2005, and the extreme rainfall events in Mumbai, India in July 2005. Could a human influence such as increased concentrations of greenhouse gases in the atmosphere have "caused" any of these events?

[INSERT FAQ 9.1, FIGURE 1 HERE]

Extreme events usually result from a combination of factors. For example, several factors contributed to the extremely hot European summer of 2003 including a persistent high pressure system that was associated with very clear skies and dry soil, which left more solar energy available to heat the land because less energy was consumed to evaporate moisture from the soil. Similarly, the formation of a hurricane requires warm sea surface temperatures and specific atmospheric circulation conditions. Because some factors may be strongly affected by human activities, such as sea surface temperatures, but others may not, it is not simple to detect a human influence on a single, specific extreme event.

Nevertheless, it may be possible to use climate models to determine whether human influences have changed the likelihood of certain types of extreme events. For example, in the case of the 2003 European heat wave, a climate model was run including only historical changes in natural factors that affect the climate, such as volcanic activity and changes in solar output. Next, the model was run again including both human and natural factors, which produced a simulation of the evolution of the European climate that was much closer to that which had actually occurred. Based on these experiments, it was estimated that over the 20th century, human influences more than doubled the risk of having a summer in Europe as hot as that of 2003, and that in the absence of human influences, the risk would probably have been one in many hundred years. More detailed modelling work will be required to estimate the change in risk for specific high impact events, such as the occurrence of a series of very warm nights in an urban area such as Paris.

The value of such a probability-based approach – "Does human influence change the likelihood of an event?" – is that it can be used to estimate the influence of external factors, such as increases in greenhouse gases, on the frequency of specific types of events, such as heat waves or frost. Nevertheless, careful statistical analyses are required, since the likelihood of individual extremes, such as a late-spring frost, could change due to changes in climate variability as well as changes in average climate conditions. Such analyses rely on climate model-based estimates of climate variability, and thus the climate models used should adequately represent that variability.

The same likelihood-based approach can be used to examine changes in the frequency of heavy rainfall or floods. Climate models predict that human influences will cause an increase in many types of extreme events, including extreme rainfall. There is already evidence that, in recent decades, extreme rainfall has increased in some regions, leading to an increase in flooding.

Frequently Asked Question 9.2: Can the Warming of the 20th Century be Explained by Natural Variability?

It is very unlikely that the 20th century warming can be explained by natural causes. The late 20th century has been unusually warm. We know from paleoclimatic reconstructions that the 2nd half of the 20th century was likely the warmest 50-year period in the Northern Hemisphere in the last 1300 years. This rapid warming is consistent with the scientific understanding of how the climate should respond to a rapid increase in greenhouse gases like that which has occurred over the past century, and the warming is inconsistent with the scientific understanding of how the climate should respond to natural external factors such as variability in solar output and volcanic activity. Climate models provide a suitable tool to study the various influences on the Earth's climate. When the effects of increasing levels of greenhouse gases are included in the models, as well as natural external factors, the models produce good simulations of the warming that has occurred over the past century. The models fail to reproduce the observed warming when run using only natural factors. When human factors are included, the models also simulate a geographic pattern of temperature change around the globe similar to that which has occurred in recent decades. This spatial pattern, which has features such as a greater warming at high northern latitudes, differs from the most important patterns of natural climate variability that are associated with internal climate processes, such as El Niño.

Variations in the Earth's climate over time are caused by natural internal processes, such as El Niño, as well as changes in external influences. These external influences can be natural in origin, such as volcanic activity and variations in solar output, or caused by human activity, such as greenhouse gas emissions, human-sourced aerosols, ozone depletion, and land use change. The role of natural internal processes can be estimated by studying observed variations in climate and by running climate models without changing any of the external factors that affect climate. The effect of external influences can be estimated with models by changing these factors, and also by using physical understanding of the processes involved. The combined effects of natural internal variability and natural external factors can also be estimated from climate information recorded in tree rings, ice cores and other types of natural "thermometers", prior to the industrial age.

The natural external factors that affect climate include volcanic activity and variations in solar output. Explosive volcanic eruptions occasionally eject large amounts of dust and sulphate aerosol high into the atmosphere, temporarily shielding the earth and reflecting sunlight back to space. Solar output has an 11-year cycle and may also have longer-term variations. Human activities over the last 100 years, particularly the burning of fossil fuels, have caused a rapid increase in carbon dioxide and other greenhouse gases in the atmosphere. Before the industrial age, these gases had remained at near stable concentrations for thousands of years. Human activities have also caused increased concentrations of fine reflective particles, or "aerosols", in the atmosphere, particularly during the 1950s and 1960s.

Although natural internal climate processes, such as El Niño, can cause variations in global mean temperature for relatively short periods, a large portion of the change in global average temperature over the 20th century has likely been caused by external factors. Brief periods of global cooling have followed major volcanic eruptions, such as Mount Pinatubo in 1991. In the early part of the 20th century, global average temperature rose, during which time greenhouse gas concentrations started to rise, solar output was probably increasing, and there was little volcanic activity. During the 1950s and 1960s, average global temperatures levelled off, as increases in aerosols from fossil fuels and other sources cooled the planet. The eruption of Mt. Agung in 1963 also put large quantities of reflective dust into the upper atmosphere. The rapid warming observed since the 1970s has occurred in a period when the increase in greenhouse gases has dominated over all other factors.

Numerous experiments have been conducted using climate models to determine the likely causes of the 20th century climate change. These experiments indicate that models cannot reproduce the rapid warming observed in recent decades when they only take into account variations in solar output and volcanic activity. However, as shown in Figure 1, models are able to simulate the observed 20th century changes in temperature when they include all of the most important external factors, including human influences from sources such as greenhouse gases, and natural external factors. The model estimated responses to these external factors are detectable in the 20th century climate globally and in each individual continent except

Antarctica, where there are insufficient observations. The human influence on climate very likely dominates over all other causes of change in global average surface temperature during the past half century.

[INSERT FAQ 9.2, FIGURE 1 HERE]

An important source of uncertainty arises from the incomplete knowledge of some external factors, such as human-sourced aerosols. Also, the climate models themselves are imperfect. Nevertheless, all models simulate a pattern of response to greenhouse gas increases from human activities that is similar to the observed pattern of change. This pattern includes more warming over land than over the oceans. This pattern of change, which differs from the principal patterns of temperature change associated with natural internal variability, such as El Niño, helps to distinguish the response to greenhouse gases from that of natural external factors. Another example is that models and observations both show warming in the lower part of the atmosphere (the troposphere), and cooling higher up in the stratosphere. This is another "fingerprint" of change that reveals the effect of human influence on the climate. If, for example, an increase in solar output had been responsible for the recent climate warming, both the troposphere and the stratosphere would have warmed. In addition, differences in the timing of the human and natural external influences help to distinguish the climate responses to these factors. Such considerations increase confidence that human, rather than natural factors were the dominant cause of the global warming observed over the last 50 years.

Estimates of Northern Hemispheric temperatures over the last one to two millennia, based on natural "thermometers" such as tree rings that vary in width or density as temperatures change, and historical weather records, provide additional evidence that the 20th century warming cannot be explained by only natural internal variability and natural external forcing factors. Confidence in these estimates is increased because prior to the industrial era, much of the variation they show in Northern Hemisphere average temperatures can be explained by episodic cooling caused by large volcanic eruptions and by changes in the sun's output. The remaining variation is generally consistent with the variability simulated by climate models in the absence of natural and human-induced external factors. While there is uncertainty in the estimates of past temperatures, they show that it is likely that the second half of the 20th century was the warmest 50-yr period in the last 1300 years. The estimated climate variability caused by natural factors is small compared to the strong 20th century warming.

Tables

Table 9.1. Inverse estimates of aerosol forcing from detection and attribution studies and studies estimating equilibrium climate sensitivity (see Section 9.6 and Table 9.3 for details on studies). The 5–95% estimates for the range of aerosol forcing relate to total or net fossil fuel-related aerosol forcing (in W m^{-2}). *First row*: source study. *Second row*: observational data used to constrain aerosol forcing. *Third row*: external forcings accounted for in the study. *Fourth row*: year(s) for which aerosol forcing is calculated, relative to pre-industrial conditions. *Fifth row*: 5–95% inverse estimate of the total aerosol forcing in the year given relative to preindustrial forcing. The aerosol range refers to the net fossil-fuel related aerosol range, which tends to be all forcings not directly accounted for that project onto the pattern associated with fossil fuel aerosols, and includes all unknown forcings and those not explicitly considered (for example, OzT and BC+OM in several of the studies). *Key to forcings*: G, greenhouse gases;. Sul, direct sulfate aerosol effect; Suli, (first) indirect sulfate aerosol effect; OzT, tropopspheric ocone; OzS, stratospheric ozone; Vol, volcanic forcing; Sol, solar forcing; BC+OM, black carbon and organic matter from fossil fuel and biomass burning.

Study	*Forest et al. (2006)*	*Andronova and Schlesinger (2001)*	*Knutti et al. (2002, 2003)*	*Gregory et al. (2002a)*	*Stott et al. (2006c)*	*Harvey et al. (2002)*
Data used	Upper air, surface and deep ocean space-time temperature, latter half of 20th century	Global mean and hemispheric difference in SAT 1856–1997	Global mean ocean heat uptake 1955–1995, mean SAT inc. 1860–2000	Surface air temperature space-time patterns, 1 AOGCM	Surface air temperature space-time patterns, 3 AOGCMs	Global mean and hemispheric difference in SAT 1856–2000
Forcings considered	G, Sul, Sol, Vol, OzS, land surface changes	G, OzT, Sul, Sol, Vol	G, OzT, OzS, fossil fuel and biomass burning BC+OM, strat. Water vapor, Vol, Sol	G, Anthrop (Sul, Suli, OxT, OzS)., natural forcings (solar volcanic)	G, Sul, Anthrop., natural forcings (solar volcanic)	G, Sul, biomass aerosol, Sol, Vol
Year	1980's	1990	2000	2000	2000	1990
Aerosol forcing [W/m^2]	–0.14 to –0.74 –0.07 to –0.65 with expert prior	–0.54 to –1.3	0 to –1.2 ind. aerosol, –0.6 to –1.7 total aerosol	–0.4 to –1.6 total aerosol	–0.4 to –1.4 total aerosol	Fossil fuel aerosol unlikely <–1, biomass+dust unlikely <–0.5[a]

Notes:
(a) Explores IPCC TAR range of climate sensitivity (i.e., 1.5-4.5ºC), while other studies explore wider ranges

Table 9.2. Components of the rate of global mean sea level rise in mm yr^{-1} from models and observations. All ranges are 5–95%. The observational components and the observed rate of sea level rise ("Obs" column) are repeated from Chapter 5, Section 5.5.6 and Table 5.3. The "ALL" column is computed (following the methods of Gregory and Huybrechts, 2006 and Chapter 10, Section 10.6.3.1) from eight 20C3M simulations that include both natural and anthropogenic forcings (model IDs 3, 9, 11, 12, 14, 15, 19 and 21; see Chapter 8, Table 8. 1), and the "ALL/ANT" column from 16 simulations: the eight ALL and eight others that have anthropogenic forcings only (model IDs 4, 6, 7, 8, 13, 16, 20 and 22; see Chapter 8, Table 8. 1).

	1961–2003			1993–2003		
	Obs	ALL	ALL/ANT	Obs	ALL	ALL/ANT
Thermal expansion	0.42 ± 0.12	0.5 ± 0.2	0.7 ± 0.4	1.60 ± 0.50	1.5 ± 0.7	1.2 ± 0.9
Glaciers and ice caps	0.50 ± 0.18	0.5 ± 0.2	0.5 ± 0.3	0.77 ± 0.22	0.7 ± 0.3	0.8 ± 0.3
Ice sheets (observed)	0.19 ± 0.43			0.41 ± 0.35		
Sum of components	1.1 ± 0.5	1.2 ± 0.5	1.4 ± 0.7	2.8 ± 0.7	2.6 ± 0.8	2.4 ± 1.0
Observed rate of rise	1.8 ± 0.5			3.1 ± 0.7		

Table 9.3. Results from key studies on observational estimates of equilibrium climate sensitivity (in °C) from instrumental data (white background), data for the last millennium (yellow), individual volcanic eruptions (purple) and simulations of the Last Glacial Maximum (LGM, blue). *First column*: the study. *Second column*: observational data used to constrain sensitivity (with range covered by uniform prior if narrower than 0–20°C). The final three rows (orange background) list some studies using non-uniform priors. *Third column*: type of model used. *Fourth column*: external forcings taken into account. *Fifth column*: uncertainties taken into account (for example, uncertainty in ocean diffusivity κ , or total aerosol forcing). Ideally, studies account for model uncertainty, forcing uncertainty (for example, aerosol and other anthropogenic forcing uncertainty ε_{aer} , and uncertainty in natural forcing ε_{nat}), uncertainty in observations, ε_{obs} , and internal climate variability ("Noise"). *Last column*: estimated equilibrium climate sensitivity. *Key to forcings*: G, greenhouse gases; Sul, direct sulphate aerosol effect; Suli, (first) indirect sulphate effect; OzT, tropospheric ozone; OzS, stratospheric ozone; Vol, volcanism; Sol, solar; BC+OM, black carbon and organic matter. EMIC numbers give the name of related EMICs described in Chapter 8, Table 8.3).

Study	Data	Model	Forcing	Treatment of uncertainties	Sensitivity Range 5–95% [°C]
Forest et al. (2006)	Upper air, surface and deep ocean space-time 20th C Temp. Prior 0-10°C	2-D EMIC (~E6)	G, Sul, Sol, Vol, OzS, land surface changes (2002: G, Sul, OzS)	ε_{obs} , noise, κ , aer , sensitivity tests for solar/volc. forcing unc.;	2.1 to 8.9 (1.4 to 7.7 without natural forcings)
Andronova and Schlesinger (2001)	Global mean and hemispheric difference in SAT 1856–1997	EBM with ocean	G (detailed), OzT, Sul, Sol, Vol, Suli	Noise (bootstrap residual), choice of radiative forcing factors	1.0 to 9.3 p>54% that α outside 1.5–4.5
Knutti et al. (2002; 2003)	Global mean ocean heat uptake 1955–1995, mean SAT 1860–2000 Prior 0-10°C	EMIC ~E1, (+ neural net)	G, OzT, OzS, fossil fuel and biomass burning BC+OM, strat. water vapour, Vol, Sol	ε_{obs} ε_{forc} for multiple forcings from IPCC (2001), κ , different ocean mixing schemes	2.2 to 9.2 p> 50% that α outside 1.5–4.5
Gregory et al. (2002a)	Global mean change in SAT and ocean heat change between (1861–1900) and (1957–1994)	1-box	G, Sul and Suli (top down via Stott et al., 2001), Sol, Vol	ε_{obs} , ε_{forc}	1.1 to ∞
Frame et al. (2005)	Global change in surface temperature	EBM	G, accounted for other anthrop and natural forcing, by fingerprints, Sul, Nat	Noise, uncertainty in amplitude but not pattern of natural and anthrop. forcings and response (scaling factors), κ (range consist. with ocean warming)	1.2 to 11.8
Forster and Gregory (2006)	1985–1996 ERBE data 60N–60S, global surface T Prior 0-18.5°C, transform after Frame et al. (2005)	1-Box	Ghg, Vol, Sol, Sul	ε_{obs} , forcing uncertainty	1.2 to 14.2
Wigley et al. (2005b)	Global mean surface temperature	Box model	From volcanic forcing only	El Nino	Agung: 1.3–6.3 El Chichon: 0.3–7.7 Pinatubo: 1.8–5.2
Hegerl et al. (2006a)	NH mean SAT preindustrial (1270/1505 to 1850) from multiple reconstructions Prior 0-10°C	EBM	G, Sul, Sol, Vol	Noise (from residual), κ , uncertainty in solar and volcanic forcing, uncertainty in reconstructions	1.1 to 8.6
Schneider von Deimling et al.	LGM tropical SSTs and other LGM data	EMIC (~E3)	LGM forcing: greenhouse gases,	uncertainty of proxy-based ice age SSTs (one	1.2 to 4.3°C (based on

(2006)			dust, ice sheets, vegetation, insolation	type of data); attempt to account for structural unc., estimate of forcing uncertainty	encompassing several ranges given)
Annan et al. (2005)	LGM tropical SSTs, Present day seasonal cycle of a number of variables for sampling prior of model parameters	AGCM with mixed layer ocean	PMIP2 LGM forcing	Obs. uncertainty in tropical SST estimates (one type of data)	<7% chance of sensitivity >6°C
Forest et al.(2002; 2006)	Expert prior, 20th century temperature change (see above)	See Forest et al.		See individual estimates	1.9-4.7
Annan et al. (2006)	Estimates from LGM, 20th century change, volcanism combined	See Annan et al.	NA	See individual estimates	1.5 to 4.5
Hegerl et al. (2006a)	1950–2000 surface temperature change (Frame et al., 2005), NH mean preindustrial SAT from last millenium	See Hegerl et al. and Frame et al.	NA	See individual estimates	1.5 to 6.2

Table 9.4. A synthesis of climate change detection results. **a)** Surface and atmospheric temperature evidence. **b)** Evidence from other variables. Note that our likelihood assessments are reduced compared to individual detection studies in order to take into account remaining uncertainties (see Section 9.1.2), such as forcing and model uncertainty not directly accounted for in the studies. The likelihood assessment is indicated in percentage terms, in parentheses where the term is not from the standard IPCC likelihood levels.

a)

Result	**Region**	**Likelihood**	**Factors contributing to likelihood assessment**
Surface temperature			
Warming during the past half century cannot be explained without external radiative forcing	Global	Extremely likely (>95%)	Anthropogenic change has been detected in surface temperature with very high significance levels (less than 1%). Conclusion is strengthened by detection of anthropogenic change in upper ocean with high significance level. Upper ocean warming argues against the surface warming being due to natural internal processes. Observed change is very large relative to climate model simulated internal variability. Surface temperature variability simulated by models is consistent with variability estimated from instrumental and paleo records. Main uncertainty from forcing and internal variability estimates (Sections 9.4.1.2, 9.4.1.4, 9.5.1.2, 9.3.4.2, 9.7).
Warming during the past half century is not solely due to known natural causes	Global	Very Likely	This warming took place at a time when non-anthropogenic external factors would likely have produced cooling. The combined effect of known sources of forcing would have been very likely to produce a warming. No climate model that has used natural forcing only has reproduced the observed global warming trend over the 2nd half of the 20th century. Main uncertainties arise from forcing, including solar, model simulated responses and internal variability estimates (Sections 9.4.1.2, 9.4.1.4, Figures 9.5, 9.6, 9.9)
Greenhouse gas forcing has been the dominant cause of the observed global warming over the last 50 years.	Global	Very likely	All multi-signal detection and attribution studies attribute more warming to greenhouse gas forcing than to a combination of all other sources considered, including internal variability, with a very high significance. Conclusion accounts for observational, model and forcing uncertainty, and the possibility that the response to solar forcing could be underestimated by models. Main uncertainty from forcing and internal variability estimates (Section 9.4.1.4, Figure 9.9).
Greenhouse gases would have caused more warming than observed over the last 50 years, with some warming offset by net cooling caused by other anthropogenic and natural forcings.	Global	Likely	Estimates from different analyses using different models show consistently more warming than observed over the last 50 years at the 5% significance level (Figure 9.4.4). However, separation of response to non-greenhouse gas (particularly aerosol) forcing from greenhouse gas forcing varies between models. (Section 9.4.1.4, Figure 9.9)

There has been a substantial anthropogenic contribution to surface temperature increases in every continent except Antartctica since the middle of the 20th century	Africa, Asia, Australia, Europe, North America and South America	Likely	Anthropogenic change has been estimated using detection and attribution methods on every individual continent (except Antarctica). Greater variability compared to other continental regions makes detection more marginal in Europe. No climate model that used natural forcing only reproduced the observed continental mean warming trend over the 2nd half of the 20th century. Uncertainties arise because sampling effects result in lower signal to noise ratio on continental than on global scales. Separation of the response to different forcings is more difficult on these spatial scales (Section 9.4.2; FAQ 9.2, Figure 1)
Early 20th century warming due in part to external forcing.	Global	Very Likely	A number of studies detect the influence of external forcings on early 20th century warming, including a warming from anthropogenic forcing. Both natural forcing and response are uncertain, and different studies find different forcings dominant. Some studies indicate that internal variability could have made a large contribution to early 20th century warming. Some observational uncertainty in early 20th century trend. (Sections 9.3.4.1, 9.4.1.4; Figures 9.4, 9.5).
Preindustrial temperatures were influenced by natural external forcing (~1300–1850)	NH (mostly extra-tropics)	Very Likely	Detection studies indicate that external forcing explains a substantial fraction of inter-decadal variability in NH temperature reconstructions. Simulations in response to estimates of preindustrial forcing reproduce broad features of reconstructions. Substantial uncertainties in reconstructions and past forcings are unlikely to lead to a spurious agreement between temperature reconstructions and forcing reconstructions as they are derived from independent proxies. (Section 9.3.4; Figures 9.4, 6.13)
Temperature extremes have changed due to anthropogenic forcing	NH land areas and Australia combined.	Likely	A range of observational evidence indicates that temperature extremes are changing. An anthropogenic influence on the temperatures of the 1, 5, 10 and 30 warmest nights, coldest days and coldest nights annually has been formally detected and attributed in one study, but observed change in the temperature of the warmest day annually is inconsistent with simulated change. The detection of changes in temperature extremes is supported by other comparisons between models and observations. Model uncertainties in changes in temperature extremes are greater than for mean temperatures and there is limited observational coverage and substantial observational uncertainty. (Section 9.4.3)
Free atmosphere changes			
Tropopause height increases detectable and attributable to anthropogenic forcing (latter half 20th century)	Global	Likely	Robust detection of anthropogenic influence on increasing tropopause height. Simulated tropopause height increases result mainly from greenhouse gas increases and stratospheric ozone decreases. Detection and attribution studies rely on reanalysis data, which are subject to inhomogeneities related to differing availability and quality of input data, although tropopause height increases have also been identified in radiosonde observations. Overall tropopause height increases in recent model and one reanalysis (ERA-40) appear to be driven by similar large scale changes in atmospheric temperature, although errors in tropospheric warming and stratospheric cooling could lead to partly spurious agreement in other datasets (Section 9.4.4.2, Figure 9.14)

Tropospheric warming detectable and attributable to anthropogenic forcing (latter half 20th century)	Global	Likely	Robust detection and attribution of anthropogenic influence on tropospheric warming, which does not depend on including stratospheric cooling in the fingerprint pattern of response. Observational uncertainties in radiosonde and satellite record. Models generally predict a relative warming of the free troposphere compared to the surface in the tropics since 1979 which is not seen in the radiosonde record (possibly due to uncertainties in the radiosonde record), but is seen in one version of the satellite record, although not others. (Section 9.4.4).
Simultaneous tropospheric warming and stratospheric cooling due to anthropogenic forcing (latter half of 20th century)	Global	Very Likely	Simultaneous warming of the troposphere and cooling of the stratosphere due to natural factors is less likely than warming of the troposphere or cooling of the stratosphere alone. Cooling of the stratosphere is in part related to decreases in stratospheric ozone. Modelled and observational uncertainties as discussed under entries for tropospheric warming with additional uncertainties due to stratospheric observing systems and the relatively poor representations of stratospheric processes and variability in climate models. (Section 9.4.4).

b)

Result	**Region**	**Likelihood**	**Factors contributing to likelihood assessment**
Ocean changes			
Anthropogenic forcing has warmed the upper several hundred meters of the ocean during latter half of the 20th century	Global (but with limited sampling in some regions)	Likely	Robust detection and attribution of anthropogenic fingerprint of ocean temperature changes in three different models, and in ocean heat content data suggests high likelihood, but observational and modelling uncertainty remains. MMD 20th century simulations simulate comparable ocean warming to observations only if anthropogenic forcing included. Simulated and observed variability appear inconsistent, either due to sampling errors in the observations or undersimulated internal variability in the models. Limited geographical coverage in some ocean basins (Section 9.5.1.2, Figure 9.15)
Anthropogenic forcing contributed to sea level rise during the latter half 20th century	Global	Very likely	Natural factors alone do not satisfactorily explain either the observed thermal expansion of the ocean or the observed sea level rise. Models including anthro-pogenic and natural forcing simulate the observed thermal expansion since 1961 reasonably well. Anthropogenic forcing dominates the surface temperature change simulated by models, and has likely contributed to the observed warming of the upper ocean and widespread glacier retreat. It is very unlikely that the warming during the past half century is due only to known natural causes. It is therefore very likely that anthropogenic forcing contributed to sea-level rise associated with ocean thermal expansion and glacier retreat. However, it remains difficult to estimate the anthropogenic contribution to sea-level rise because suitable studies quantifying the anthropogenic contribution to sea level rise and glacier retreat are not available, and because the observed sea level rise budget is not closed (see Table 9.2; Section 9.5.2).

Circulation			
Sea level pressure shows detectable anthropogenic signature during latter half 20th century	Global	Likely	Changes of similar nature observed in both hemispheres and are qualitatively, but not quantitatively consistent with model simulations. Uncertainty in models and observations. Models underestimate the observed NH changes for reasons that are not understood, based on a small number of studies. Simulated SH response to ozone changes is consistent with observations (Section 9.5.2.4, Figure 9.16).
Anthropogenic forcing contributed to increase in tropical cyclone intensity since the 1970s	Tropical regions	More likely than not (>50%)	Recent observational evidence suggests an increase in frequency of intense storms. Increase in intensity is consistent with (but stronger than) theoretical expectations. Large uncertainties due to models and observations. Process and model studies anticipate only modest changes in maximum intensity by the end of the 21st century. Observational evidence, which is affected by substantial inhomogeneities in tropical cyclone data sets for which corrections have been attempted, suggests that increases in cyclone intensity since the 1970s are associated with SST and atmospheric water vapour increases. (Section 3.8.3, Box 3.4 and Section 9.5.2.6)
Precipitation, Drought, Runoff			
Volcanic forcing influences total rainfall	Global land areas	More likely than not (>50%)	Model response detectable in observations for some models and result supported by theoretical understanding. However, uncertainties in models, forcings and observations. Limited observational sampling, particularly in the Southern Hemisphere (Section 9.5.3.2, Figure 9.18).
Increases in heavy rainfall consistent with anthropogenic forcing during latter half 20th century	Global land areas (limited sampling)	More likely than not (>50%)	Observed increases in heavy precipitation appear to be consistent with expectations of response to anthropogenic forcing. Models may not represent heavy rainfall well; observations suffer from sampling inadequacies. (Section 9.5.3.2)
Increased risk of drought due to anthropogenic forcing during latter half 20th century	Global land areas	More likely than not (>50%)	One detection study has identified an anthropogenic fingerprint in a global PDSI (Palmer Drought Severity Index) data set with high significance, but the simulated response to anthropogenic and natural forcing combined is weaker than observed, and the model appears to have less inter-decadal variability than observed. Studies of some regions indicate that droughts in those regions are linked either to SST changes that, in some instances, may be linked to anthropogenic aerosol forcing (e.g., Sahel) or a circulation response to anthropogenic forcing (e.g., SW Australia). Models, observations and forcing all contribute uncertainty. (Section 9.5.3.2)
Cryosphere			
Anthropogenic forcing has contributed to reductions in NH sea-ice extent during latter half of 20th century	Arctic	Likely	The observed change is qualitatively consistent with model simulated changes for most models and expectation of sea ice melting under Arctic warming. Sea ice extent change detected in one study. The model used has some deficiencies in Arctic sea ice annual cycle and extent. The conclusion is supported by physical expectation and simulations with another climate model. Change in SH sea ice probably within range explained by internal variability (Section 9.5.4.1).
Anthropogenic forcing has contributed to widespread glacier retreat during 20th century	Global	Likely	Observed changes qualitatively consistent with theoretical expectations and temperature detection. Anthropogenic contribution to volume change difficult to estimate. Few detection and attribution studies, but retreat in vast majority of glaciers consistent with expected reaction to widespread warming. (Section 9.5.4.3)

Appendix 9.A: Methods Used to Detect Externally Forced Signals

We very briefly review the statistical methods that have been used in most recent detection and attribution work. Standard 'frequentist' methods (methods based on the relative frequency concept of probability) are most often used, but there is also increasing use of Bayesian methods of statistical inference. We will briefly describe the optimal fingerprinting technique in the following subsection. This will be followed by a short discussion on the differences between the standard and Bayesian approaches to statistical inferences that are relevant to detection and attribution.

9.A.1 Optimal Fingerprinting

Optimal fingerprinting is generalized multivariate regression adapted to the detection of climate change and the attribution of change to externally-forced climate change signals (Hasselmann, 1979, 1997; Allen and Tett, 1999). The regression model has the form $\mathbf{y} = \mathbf{Xa} + \mathbf{u}$, where vector $\mathbf{y}$ is a filtered version of the observed record, matrix $\mathbf{X}$ contains the estimated response patterns to the external forcings (signals) that are under investigation, $\mathbf{a}$ is a vector of scaling factors that adjusts the amplitudes of those patterns, and $\mathbf{u}$ represents internal climate variability. Vector $\mathbf{u}$ is usually assumed to be a Gaussian random vector with covariance matrix $\mathbf{C}$. Vector $\mathbf{a}$ is estimated with $\hat{\mathbf{a}} = (\mathbf{X}^T\mathbf{C}^{-1}\mathbf{X})^{-1}\mathbf{X}^T\mathbf{C}^{-1}\mathbf{y}$, which is equivalent to $(\widetilde{\mathbf{X}}^T\widetilde{\mathbf{X}})^{-1}\widetilde{\mathbf{X}}^T\widetilde{\mathbf{y}}$, where matrix $\widetilde{\mathbf{X}}$ and vector $\widetilde{\mathbf{y}}$ represent the signal patterns and observations after normalization by the climate's internal variability. This normalization, standard in linear regression, is used in most detection and attribution approaches to improve the signal-to-noise ratio (see e.g., Mitchell et al., 2001, Hasselmann, 1979; Allen and Tett, 1999).

The matrix $\mathbf{X}$ typically contains signals that are estimated with either an AOGCM, an atmospheric general circulation model (AGCM; see Sexton et al., 2001; Sexton et al., 2003) or a simplified climate model such as an energy balance model (EBM). Because AOGCMs simulate natural internal variability as well as the response to specified anomalous external forcing, AOGCM simulated climate signals are typically estimated by averaging across an ensemble of simulations (for a discussion of optimal ensemble size and composition, see Sexton et al., 2003). If an observed response is to be attributed to anthropogenic influence, $\mathbf{X}$ should at a minimum contain separate natural and anthropogenic responses. In order to relax the assumption that the relative magnitudes of the responses to individual forcings are correctly simulated, $\mathbf{X}$ may contain separate responses to all the main forcings, including greenhouse gases, sulphate aerosol, solar irradiance changes and volcanic aerosol. The vector $\mathbf{a}$ accounts for possible errors in the amplitude of the external forcing and the amplitude of the climate model's response by scaling the signal patterns to best match the observations.

Fitting the regression model requires an estimate of the covariance matrix $\mathbf{C}$ (i.e., the internal variability) which is usually obtained from unforced variation simulated by AOGCMs (e.g., from long control simulations) because the instrumental record is too short to provide a reliable estimate and may be affected by external forcing. AOGCMs may not simulate natural internal climate variability accurately, particularly on small spatial scales, and thus a residual consistency test (Allen and Tett, 1999) is typically used to assess the model simulated variability on the scales that are retained in the analysis. To avoid bias (Hegerl et al., 1996; Hegerl et al., 1997), uncertainty of the estimate of the vector of scaling factors $\mathbf{a}$ is usually assessed with a second, statistically independent estimate of the covariance matrix $\mathbf{C}$ which is ordinarily obtained from an additional, independent sample of simulated unforced variation.

Signal estimates are obtained by averaging across an ensemble of forced climate change simulations, but contain remnants of the climate's natural internal variability because the ensembles are finite. When ensembles are small or signals weak, these remnants may bias ordinary least squares estimates of $\mathbf{a}$ downward. This is avoided by estimating $\mathbf{a}$ with the total least squares algorithm (Allen and Stott 2003).

9.A.2 Methods of Inference

Detection and attribution questions are assessed through a combination of physical reasoning (to determine, for example, by assessing consistency of possible responses, whether other mechanisms of change not included in the climate model could plausibly explain the observed change) and by evaluating specific hypotheses on the scaling factors contained in $\mathbf{a}$. Most studies evaluate these hypotheses using standard frequentist methods (Hasselmann, 1979, 1997; Hegerl et al., 1997; Allen and Tett, 1999). Several recent studies have also used Bayesian methods (Hasselmann, 1998; Leroy, 1998; Min et al., 2004; Lee et al., 2005; Min et al., 2005; Schnur and Hasselmann, 2005; Lee et al., 2006; Min and Hense, 2006a).

In the standard approach, detection of a postulated climate change signal occurs when its amplitude in observations is shown to be significantly different from zero (i.e., when the null hypothesis $H_D : \mathbf{a} = \mathbf{0}$ where $\mathbf{0}$ is a vector of zeros, is rejected) with departure from zero in the physically plausible direction. Subsequently, the second attribution requirement (consistency with a combination of external forcings and natural internal variability) is assessed with the *attribution consistency test* (Hasselmann, 1997; see also Allen and Tett, 1999) that evaluates the null hypothesis $H_A : \mathbf{a} = \mathbf{1}$ where $\mathbf{1}$ denotes a vector of units. This test does not constitute a complete attribution assessment, but contributes important evidence to such assessments, see Mitchell et al. (2001). Attribution studies usually also test whether the response to a key forcing, such as greenhouse gas increases, is distinguishable from that to other forcings, usually based on the results of multiple regression (see above) using the most important forcings simultaneously in $\mathbf{X}$. If the response to a key forcing, for example, due to greenhouse gas increases, is detected by rejecting the hypothesis that its amplitude $a_{GHG} = 0$ in such a multiple regression, this provides strong attribution information. The reason is that it demonstrates that the observed climate change is "not consistent with alternative, physically-plausible explanations of recent climate change that exclude important elements of the given combination of forcings" (Mitchell et al., 2001).

Bayesian approaches are of interest because they can be used to integrate information from multiple lines of evidence, and can incorporate independent prior information into the analysis. Essentially two approaches (described below) have been taken to date. In both cases inferences are based on a posterior distribution that blends evidence from the observations with the independent prior information, which may include information on the uncertainty of external forcing estimates, climate models, and their responses to forcing. In this way, all information that enters into the analysis is declared explicitly.

Schnur and Hasselmann (2005) approach the problem by developing a filtering technique that optimizes the impact of the data on the prior in a manner similar to the way in which optimal fingerprints maximize the ratio of the anthropogenic signal to natural variability noise in the conventional approach. The optimal filter in the Bayesian approach depends on the properties of both the natural climate variability and model errors. Inferences are made by comparing evidence, as measured by Bayes Factors (Kass and Raftery, 1995), for competing hypotheses. Other studies using similar approaches include Min et al. (2004) and Min and Hense (2006a). In contrast, Berliner et al. (2000) and Lee et al. (2005) use Bayesian methods only to make inferences about the estimate of $\mathbf{a}$ that is obtained from conventional optimal fingerprinting.