Chapter 9: Understanding and Attributing Climate Change

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Executive Summary

Evidence of the effect of external influences on the climate system has continued to accumulate since the TAR. There is widespread evidence of anthropogenic warming of the climate system in temperature observations taken at the surface, in the free atmospheric and in the oceans. It is very likely that greenhouse gas forcing has been the dominant cause of the observed global warming over the last 50 years, and it is likely that greenhouse gases alone would have caused more warming than has been observed during this period, with some warming offset by cooling from natural and other anthropogenic factors. It is highly likely (>95%) that warming during the past half century cannot be explained without external forcing, or based on known natural external causes alone. The warming occurred in both oceans and atmosphere, and took place at a time when non-anthropogenic external forcing factors would likely have produced cooling. Studies that attribute warming to greenhouse gas forcing rely on a range of observational data sets, variables, climate models and methods. Anthropogenic warming has now been detected on continental and sub-continental scales in a range of studies, and evidence is emerging that surface temperature extremes have likely been affected. In addition, anthropogenic influence is becoming apparent in other parts of the climate system. While many uncertainties remain, the physical consistency of the many different lines of evidence increases our confidence in the conclusion that anthropogenic forcing has caused substantial change in the climate system over the past century.

Human-induced warming of the climate system is wide-spread. Anthropogenic warming of the climate system can be detected in temperature observations taken at the surface, in the free atmosphere 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 during the past half century would likely have resulted in greater warming than observed if there had not been an offsetting cooling effect from aerosol and other forcings. This conclusion takes into account observational and forcing uncertainty, and is 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 since wide-spread radiosonde measurements started in the late 1950s. 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 observed warming of 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 and ice sheet melt, is likely the largest contributor to sea level rise during the latter half 20th century.

Anthropogenic influence is now detectable in regional surface temperatures. The anthropogenic signal in surface temperature changes has now likely been detected in all inhabited continents and many sub-continental land areas. The ability of models to simulate the temperature evolution on these scales and the detection of anthropogenic effects on individual continents provides stronger evidence of human influence on the global climate than was available to the TAR.

Evidence of anthropogenic influence on surface temperature extremes is emerging. Surface temperature extremes have likely been affected by anthropogenic forcing. Many indicators of impact-relevant surface temperatures, including the annual numbers of frost days, warm days and cold days, show changes consistent with warming. 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.

Anthropogenic influence is becoming apparent in other parts of the climate system. Anthropogenic forcing has likely contributed to recent decreases in Arctic sea ice extent. There is evidence of a decreasing trend in global snow cover and widespread retreat of glaciers, consistent with warming, and evidence that this melting has also likely contributed to sea-level rise.

Trends in the Northern and Southern Annular Modes over recent decades, which correspond to sea level pressure reductions over the poles, are likely related in part to human activity. 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.

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. Observed, large scale changes in land precipitation over the 20th century appear to be qualitatively consistent simulations of the 20th century, suggestive of a possible human influence. Observed increases in heavy precipitation appear to be consistent with increases that are expected to occur with warming, although it is unclear whether these changes are distinguishable from natural variability. The observed increase in the proportion of very intense hurricanes is the same direction but not magnitude as suggested by theoretical studies and modelling stuidies of projected 21st century change. Inadequacies in process knowledge and understanding of natural variability, modelling, and monitoring of tropical cyclones inhibit attribution to anthropogenic factors.

Proxy climate data have been used to increase confidence in models and the role of external influences on climate. Coupled climate models used to predict future climate have been used to understand past climatic conditions such as those of the Last Glacial Maximum and the Mid-Holocene. While many aspects of these past climates are still uncertain, climate models are successful in reproducing their broad features when forced with boundary conditions and radiative forcing for those periods. A large fraction of Northern Hemisphere interdecadal temperature variability in reconstructions of the past 7 centuries is very likely attributable to natural external forcing. Furthermore, the 20th century warming evident in these records, most of which end by the mid-20th century, is likely due in part to anthropogenic forcing.

Estimates of the climate sensitivity are now better constrained by observations. Estimates of the equilibrium climate sensitivity and the transient climate response, their uncertainties and the ability of observable quantities to constrain them are now better understood. The equilibrium climate sensitivity is likely 2 to 4.5°C, with a most likely value of approximately 3°C, although the upper limit remains difficult to constrain from observations. The transient climate response is better constrained and is unlikely to be greater than 2.8°C at the time of CO_2 doubling in response to a 1% per year increase in CO_2.

Overall consistency of evidence. Although uncertainties remain, many observed changes in the climate over the 20th century are distinct from internal variability and consistent with the expected response to anthropogenic forcing. This consistency is apparent in many observations, including surface and free atmospheric temperature, ocean temperature and sea-ice extent, some large scale features of the atmospheric circulation. The simultaneous increase in energy content of all the major components of the climate system supports the conclusion that the cause of the warming is highly 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 improved models and analysis techniques have 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 inter-decadal variations in solar forcing. However, 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. The extent to which uncertainties in observations, forcing and model formulation contribute to a lack of quantitative agreement is not presently understood.

The detection and attribution of external influences is performed by comparing observed changes against those that could plausibly arise from natural internal variability. Improvements in models and instrumental and proxy climate records have increased confidence in climate model simulated internal variability. However, uncertainty remains because the available observational records are influenced by external forcing, are often too short to provide accurate estimates of decadal and longer time scale variability, and some have inadequate spatial coverage and accuracy. While reduced, 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 model uncertainties still restrict detection and attribution of changes in extremes, although understanding of changes in the intensity, frequency and risk of extremes is emerging.

9.1 Introduction

The objective of this chapter is to understand the observed climate changes that are reported in Chapters 3 to 5 as expressions of natural internal climate variability and/or externally forced climate change. The scope of this chapter is wider than that of the "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 include research on regional scales, extremes and variables other than temperature not previously available. We attempt to place that work in the context of a broader understanding of a changing climate. However, the ability to interpret some changes, particularly for non-temperature variables, is sometimes limited by uncertainties in the observations, physical understanding of the climate system, available climate models, and external forcing estimates.

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 the 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 external forcings and how they may be distinguished from change and variability that results from internal climate system processes.

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, which are based on physical understanding of the climate system, are often quantified with climate models that are run with prescribed changes in external forcing. The assessment of observed changes against those that are expected is performed in a number of ways. The formal detection and attribution method that is often used (see Section 9.1.2) uses objective statistical tests to assess whether observations contain evidence of the expected responses to external forcing that are distinct from internal variability.

In order to understand climate change we must therefore also understand natural climate variability, which is composed of *internal variability* that results from internal climate processes (see Glossary) and the climate's response to natural external forcing. 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., the triggering of convection) up to years (e.g., tropospheric-stratospheric 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 of decades to centuries. These components produce internal variability directly and by integrating 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). The climate's internal variability is difficult to estimate because all climate observations are influenced, at least to some extent, by variations in external forcing. However estimates can be obtained from observations or models under certain conditions.

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" (see Glossary). In this chapter, the methods used to identify change in observations are based on the expected responses to external forcing (see 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 occurance by random chance by 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 anthropogenic influence given the expectation that anthropogenic forcing would result in moistening in some latitudes that is partially offset by drying elsewhere (see Chapter 10; see also Section 9.5.3.2). The detection of an effect of external forcing on the climate does not necessarily imply that it has an important impact on the environment, biota, or human society.

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, which are commonly referred to as *fingerprints*, are usually derived from changes simulated by a climate model in response to forcing. The complexity of such a model can vary from one that is based on simple physical concepts (such as energy balance models) to complex coupled atmosphere-ocean general circulation models. 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, in many regions precipitation and temperature are ordinarily inversely correlated, with increases in temperature corresponding to drying conditions. Thus a warming trend in a given region that is not associated with rainfall change may indicate an external influence on the climate of that region (Nicholls et al., 2005) (see Section 9.4.2.3).

The spatial and temporal scales used to analyze climate change are carefully chosen so as to focus on the space-time scale of the response, filter out internal variability 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 over time, and thus analyses often smooth data to remove small-scale variations. Similarly, when fingerprints are used, averaging over an ensemble of coupled model simulations helps to isolate the model's fingerprint (i.e., response to forcing) from model simulated internal variability.

Because detection studies are necessarily statistical in nature, the inferences that can be made about whether an external influence has been detected can never be absolutely certain. It is always possible that a significant result at, say, the 5% level, could simply reflect a rare event that would have occurred in any case with less than 1 chance in 20 in an unchanged climate. Corroborating lines of evidence providing a physically consistent view of the likely cause for the changes reduce the risk of such spurious detection.

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 pertinent. 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 for such questions, and examples of studies that attempt to do so are discussed in this chapter. A promising approach, which has now been applied in at least one study, is to use information from both models and observations combined to estimate the impact of external forcing on the likelihood of specific types of rare events.

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. That, of course, is not possible, and thus from a practical perspective, attribution of anthropogenic climate change is understood to mean (i) detection, (ii) demonstration that the detected change is "consistent with the estimated responses to the given combination of anthropogenic and natural forcing", and (iii) demonstration that the detected change is "not consistent with alternative, physically-plausible explanations of recent climate change that exclude important elements of the given combination of forcings" (IPCC, 2001).

The second requirement, the assessment of the consistency between an observed change and the estimated response to a hypothesized forcing, is often achieved by determining whether the amplitude of the hypothesized pattern of change estimated from observations is statistically consistent with expectations (see

Appendix 9.A). If so, the evidence for a causal connection is substantially increased. Note however, that this statistical consistency forms only a part of the evidence that is used in attribution studies. Another key element is the consideration of the physical consistency of multiple lines of evidence.

The third requirement is to evaluate the possibility that the observed change is consistent with alternative explanations that exclude important elements of a given combination of forcings that are hypothesized to have influenced the climate. Physical understanding plays an important role in such an evaluation, but statistical analysis that identifies the separate influences of the individual forcing agents in observations is also important (see Appendix 9.A). 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 (see Section 9.2).

All three aspects of 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 short relative to the timescales of interest, and because of uncertainties in the forcings and the estimated responses. Thus internal climate variability is also 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. Confidence is further increased by systematic intercomparison of the ability of models to simulate the various modes of observed variability (see Chapter 8), by comparisons between variability in observations and climate model data (see Section 9.4), and by comparison between proxy reconstructions and climate simulations of the last millennium (see Chapter 6 and Section 9.3).

Results where attribution is not achieved, for example, where an expected pattern of change is detected, but with an amplitude that 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.

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, Chapter 10), and in the representation of physical processes in models (structural uncertainty). Such an assessment is not yet available, although research with that goal in mind is underway and model intercomparison studies (see Chapter 8) continue to improve our appreciation of these uncertainties. The effects of forcing uncertainties, which can be considerable for some forcing agents, such as solar and aerosol forcing (see Section 9.2), also remain difficult to evaluate, despite advances in research. Detection and attribution results that are based on several models or several forcing histories do provide information on the effects of model and forcing uncertainty that leads towards a more reliable attribution of climate change to a cause. Such results 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.

Full detection and attribution studies are not yet feasible for all variables for a variety of reasons. This is particularly the case for variables such as rainfall that are less reliably modelled or observed, or respond less strongly to external forcing. 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 our understanding of climate change and is therefore discussed in this chapter.

The approaches used in detection and attribution research described above can not fully account for all uncertainties, and thus ultimately expert judgement is used to estimate the likelihood that a specific cause is responsible for a given climate change. The approach in making these judgments used in this chapter is to assess results from multiple studies using a variety of models, forcings, analysis techniques, and observational data sets and to subsequently conservatively assess the likelihood of the hypothesized link at a likelihood level below the consensus level of significance inferred by the different studies. Such a decrease in significance levels by expert judgement attempts to account for remaining uncertainties that are not accounted for, such as limited range of exploration of possible forcing histories of uncertain forcings, or structural uncertainties.

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 2 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 (SAR; 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, output from several climate models for fingerprints and estimates of internal climate variability, 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 simulated greenhouse gas response was generally found to be consistent with the observed greenhouse response 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 climate's response to external forcing) as well as assumptions made and techniques used in detection and attribution studies.

The TAR also reported on a wide range of evidence of qualitative consistencies between observed climate changes and model responses to anthropogenic forcing, including global warming, 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 from models and observations. 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 were also reported in our knowledge of external forcing, particularly in anthropogenic aerosol forcing, and reconstructions of 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 20^{th} century warming by natural forcings and anthropogenic forcings other than greenhouse gases showed some discrepancies with model simulations and were model dependent. These results made it difficult to attribute the observed climate change to one specific combination of external influences.

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".

9.1.4 Scope of this Chapter

We conclude this section by briefly by providing the reader with a road map of the following sections. The body of this chapter starts with a description of forcing and its uncertainty during the recent history of the climate system, and a brief summary of the climate response associated with these forcings (see Section 9.2). The description of the response focuses on the similarities, differences and uncertainties in the climate's response to forcing because this has important implications on the attribution of observed changes to causes. The description of forcings also includes an assessment of studies that use so-called "inverse" methods. Such methods use the observed climate change to make inferences about the likely magnitude of uncertain radiative forcings. Overall, this section attempts a synthesis of results from Chapters 2 and 8 of this report that are relevant for this chapter.

Section 9.3 follows with a synthesis of paleo-climate material that draws very heavily on Chapter 6 in order to gain further insight into the link between external forcing and climate. This section confines itself to three periods in recent climate history (the last millennium, the time of the last glacial maximum at 21,000 years before present, and a period in the Holocene that is approximately 6,000 years before present) where the forcing situation is relatively well understood, and for which there are relatively large amounts of proxy data.

Section 9.4 focuses on understanding instrumental era temperature change at the surface and in the free atmosphere. This section, which draws on observed changes described in Chapters 3 and 8, reports recent developments in the detection and attribution of these changes to external forcing changes on global, continental and regional scales. It also describes an emerging body of work on the links between external forcing and the possibility of change in the intensity and/or frequency of extreme temperature events.

Section 9.5 assesses large scale climate change in other variables and climate components. This section draws on observed changes described in Chapters 3, 4, 5 and 8. Where possible, it attempts to attribute these changes to external forcing or to identify links between related changes, such as those linking some aspects of sea-surface temperature change with precipitation change

Section 9.6 completes the chapter with an assessment of studies that use comparisons between model simulations and instrumental observations or paleoclimatic data to constrain estimates of key climate parameters such as its equilibrium climate sensitivity.

9.2 Radiative Forcing and Climate Response

There are two basic types of calculations used in detection and attribution studies. The first uses the modelers' best estimate of forcing together with their best estimate of modelled climate processes to calculate the effects of external changes in the climate system (forcings) on the change in climate. These so-called "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 responses to the forcings. These forward calculations are explored further in Section 9.4.

In the second type of calculation, the so-called "inverse" calculations, adjustments to the magnitude of one or more uncertain aspects of the forward model (including the forcing that is applied) are determined in order to provide a best fit to the observational record. Such inverse methods can account for both uncertainties in different aspects of the forward model as well as uncertainties in the observations and in forcing. In general, the greater the degree of a priori uncertainty in some aspect of the model, the greater the model is allowed to adjust. Probabilistic estimates for model parameters and the forcings explored are obtained by comparing the agreement between simulations and observations, and taking into account prior uncertainties (see Sections 9.2.1.2 and 9.6).

Results from forward calculations are often further interpreted by means of a formal detection and attribution analysis. In such studies the forward model is used to calculate a response pattern for a particular forcing or set of forcings (i.e., a “fingerprint”). A linear combination of these response patterns is then adjusted (scaled) to provide the best fit to the observations. Errors or uncertainties in the magnitude of the forcing or the magnitude of a model’s response to the forcing should not affect these 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 scaling should be consistent with unity when the forcing estimates used in the forward model are consistent with those based on our understanding of the forcing (see Chapter 2). For detection studies, if the space-time pattern of response is incorrect, then the scaling, and with it detection and attribution results will be further affected.

The estimated ranges of forcing are reviewed in this section, primarily for the period since 1750, with a brief reference to forcing in 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. The response of models to different forcings, forcing uncertainties and the implications for simulated responses and climate change detection and attribution is also discussed.

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

Radiative forcing is defined as “the change in net (downward minus upward) irradiance (solar plus long-wave; in W m^{-2}) at the tropopause after allowing for stratospheric temperatures to readjust to radiative equilibrium, but with the state of the surface and tropospheric temperatures held fixed at the unperturbed values” (see Chapter 2). Estimates of the radiative forcing from forward model calculations and observations since 1750 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), stratospheric ozone decrease, tropospheric ozone increase, 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, variability in solar forcing, and stratospheric and tropospheric H_2O increase from CH_4 and irrigation. An example of one model’s response to a set of forcings is given in Chapter 2, Figure 2.26. While several of the IPCC AR4 models have included many of these forcings for the purpose of simulating the 20th century climate, most detection studies to date have used model runs with a more limited set of forcings. The total net forcing from the estimates in Chapter 2 is approximately 1.5 ± 1 W m^{-2} (67% confidence interval). As noted in Chapter 2, the estimated temperature response from these forcings is only approximately linear (i.e., the “efficacy” of each forcing, its effectiveness at changing the surface temperature, compared to carbon dioxide, is not necessarily equal to one), so that summing these fluxes does not necessarily give an adequate measure of the global average surface temperature change. Nevertheless, it is the large-scale spatial-temporal pattern of the temperature response to these forcings that is most important in detection and attribution studies rather than the specific sensitivity of the model and forcing values used.

9.2.1.2 Summary of “inverse” estimates of net aerosol forcing

Forward model approaches to estimating aerosol forcing are based on estimates of aerosol physics and chemistry and, as such, resolve the separate contributions by various aerosol components and forcing mechanisms. In contrast, inverse calculations (see Section 9.6 and Appendix 9.B for details) infer only the net forcing required to match climate model simulations with observations. The “net” forcing includes all forcings that project on the fingerprint of the forcing that is estimated. An example could be tropospheric ozone forcing projecting onto sulphate aerosol forcing. These methods can increase the ability to distinguish between responses to different external forcings by using not only a global average forcing and response, but also the spatial and temporal patterns of climate response. Inverse methods have been used to constrain both climate sensitivity and one or more forcings as well as 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 9.2.1; Stott et al., 2006c). The reliability of the spatial and temporal patterns used are discussed in Sections 9.2.2.1 and 9.2.2.2.

Forward calculations may yield a total net radiative forcing over the 20th century that could be close to zero or even negative (Boucher and Haywood, 2001). If the efficacy of the aerosol forcing is close to one, then a net forcing close to zero may allow a high value of the climate sensitivity to be consistent with the warming trend of the 20th century. However, net negative forcing would be impossible to reconcile with instrumental observations because explanations for the observed warming due to the redistribution of heat by natural internal variability are effectively ruled out by both the observed ocean warming and the natural variability in unforced climate models (or the amount of variability in paleo reconstructions that is not explained by external forcing) (Mitchell et al., 2001; see also Sections 9.3.4, 9.4.1.3, 9.5.1 and 9.7).

Since inverse calculations only yield the “net forcing”, which includes all forcings that project on the fingerprint of the forcing that is estimated, differences between forward estimates and inverse estimates may have one of several causes: (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, a possibility this is not yet ruled out since understanding and modelling of the ice phase in climate models is still in its infancy (Penner et al., 2001), or (3) other forcings project on the fingerprint of the forcing that is estimated. An example could be tropospheric ozone forcing projecting onto sulphate aerosol forcing.

Different inverse approaches consider different external forcings as is summarized in Table 9.2.1. One type of inverse method is to determine the magnitude of the response to different external forcing agents from ranges of climate change fingerprint scaling factors derived from detection analyses (Stott et al., 2000; Gregory et al., 2002a, see also Section 9.4.1.4). These effectively give the range of fingerprint magnitudes (for example, for the combined temperature response to different aerosol forcings) that are consistent with observed climate change. Scaling factors on a coupled climate model’s response to well-mixed greenhouse gases and its response to aerosol forcing can be used to infer the likely range of the combined aerosol forcing that is consistent with the observed record. This 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; Allen et al., 2005; Table 9.2.1). By scaling spatio-temporal patterns of response up or down, this technique takes account of gross errors in models but does not fully account for modelling uncertainty in patterns of temperature response to uncertain forcings.

[INSERT TABLE 9.2.1 HERE]

A more involved approach uses the response of climate models, most often simple climate models or Earth System Models of Intermediate Complexity (see Chapter 8, Table 8.8.2**)** to explore the range of forcings and climate parameters that yield results consistent with observations. Like detection methods, these approaches assess the fit of space-time patterns, or spatial means in time, to observed surface, atmospheric or ocean temperatures. They then assess the probability of combinations of climate sensitivity and net aerosol forcing based on the fit between simulations and observations (see Section 9.6 and Appendix 9.B for further discussion). These are often based on Bayesian approaches (Appendix 9.B), where prior assumptions for ranges of external forcing (frequently from forward approaches) are updated using observed climate change to obtain posterior distributions of external forcing magnitude consistent with observed climate change. However, this approach has some caveats since the spatial distribution of response in these models may be highly parameterized. The resulting range is generally narrower than that from forward-model approaches due to the additional information used (Anderson et al., 2003). However, the forward and inverse approaches are not fully comparable because the former estimate forcings that result from particular mechanisms, while the latter estimate net forcing for all mechanisms that have response patterns similar to that assumed a priori.

Studies attempt to separate greenhouse gas and aerosol effects in observations by making use of either the hemispheric gradient in forcing, or the relative difference in the temporal variation in aerosol and greenhouse gas forcing. 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 temperature can provide a useful constraint on the net aerosol radiative forcing (as found, for example, by Harvey and Kaufmann, 2002).

The ensemble-based studies summarized in Table 9.2.1 suggest that to be consistent with observed warming, the net aerosol forcing over the 20th century should be negative and with a magnitude less than about 1.7 W/m^2. Harvey and Kaufmann (2002), who use an approach that focuses on the traditional IPCC range of climate sensitivities, further conclude that global mean forcing from fossil fuel-related aerosols is "unlikely" to have exceeded –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. Results from inverse approaches (Table 9.2.1) suggest that total aerosol forcing may be somewhat weaker than the best estimates given in Table 2.11.

These estimates are affected by various uncertainties. For example, some studies use the difference between Northern and Southern Hemisphere mean temperature to separate the greenhouse gas and aerosol forcing effects (e.g., Andronova and Schlesinger, 2001; Harvey and Kaufmann, 2002). However, the ratio of Northern to Southern Hemisphere indirect forcing by industrial aerosols is not accurately known (see Chapter 2 and Section 9.2.2.2). Also, it is necessary to account for hemispheric asymmetry in tropospheric ozone forcing. Additionally, aerosols from biomass burning could cause an important fraction of the total aerosol forcing although this forcing shows little hemispheric asymmetry. Overall, results will be only as good as the spatial or time pattern that is assumed in the analysis, and missing forcings or lack of knowledge about uncertainties may hamper the interpretation of results. Nevertheless, inverse results do provide valuable additional information on the possible range of forcings if interpreted carefully taking into account the various assumptions and uncertainties.

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) (see Section 9.3) and used in estimates of climate sensitivity based on paleoclimatic records (see 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 with confidence from astronomical laws (Berger, 1978), force climate variations by changing the seasonal and latitudinal distribution of solar insolation (see Chapter 6).

Solar 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). Thus most modelling studies of this period do not treat ice-sheet extent and elevation, and CO_2 concentration prognostically, but rather specify them as boundary conditions. LGM radiative perturbation from the reduced atmospheric CO_2 concentration is estimated to have been –1.9 to –2.6 W/m^2 (best guess –2.2 W/m^2) relative to the preindustrial (Chapter 6) . Ice-sheet albedo forcing is estimated to have been –2 to –4.3 W/m^2 (Chapter 6; based on several LGM simulations) and radiative forcing from increased atmospheric aerosols (dust primarily) is estimated to have been about –1 ± 0.5 W/m^2 (see Chapter 6). Therefore, the total radiative perturbation during the LGM is estimated to have been approximately –6.1 to –10.6 W m^{-2} relative to pre-industrial with large seasonal and geographical variations (see Chapter 6).

In constrast, the major forcing difference between the preindustrial and mid-Holocene climates is due to orbital forcing, which leads to a 5% increase in summer insolation in the northern hemisphere (NH) compared to the present. The NH mean seasonal forcing was about 27 W/m^2 larger, whereas there was only a negligible change in NH annual solar forcing. For the Southern Hemisphere, there was a net annual forcing of –1 W/m^2 and for the globe, the annual 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. They led to a reduction of summer insolation of 0.33 W/m^2 at 45°N, an increase in winter insolation of 0.83 W/m^2 (Goosse et al., 2005), and a decrease in the magnitude of the insolation mean seasonal cycle of 0.4 W/m^2 in the NH between the beginning and end of the millenium. 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). For example, from approximately 1675-1715, sunspots were generally missing (the so called Maunder Minimum). Therefore, solar irradiance is believed to have been smaller than before and after this period. The estimated difference between present day solar irradiance and the late Maunder Minimum is –1.1 W/m^2 (best estimate, range –0.5 to –2 W/m^2 (see Chapter 2, 1675–1715 minus present day radiative forcing), but with large uncertainties.

Natural external forcing also results from explosive volcanism that introduces aerosols into the stratosphere (Robock and Free, 1995; IPCC, 2001), leading to a global forcing of a few W/m2 (depending on the strength of the eruption) during the year following the eruption. Several reconstructions are available for the last two millennia and have been used to force climate models (see Chapter 6). There is close agreement on the timing of large eruptions in the various compilations of historic volcanic activity, but uncertainty on the order of 50% for the size of individual eruptions (Crowley, 2000, updated; Robertson et al., 2001; for a comparison see Zorita et al., 2004; see Chapter 6), and on the order of about 40% for the overall amplitude of volcanic forcing (Crowley, 2000, updated). Different reconstructions identify similar periods when eruptions happened more frequently. These changes in the frequency of eruptions may have caused considerable inter-decadal climate variability during the past millennium (see Section 9.3.4).

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 depends on the extent to which those responses are distinct (see, for example, Section 9.4.1.4 and Appendix 9.A). Figure 9.2.1 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 9.2.2 illustrates the zonal average temperature response in the CSIRO model (with a coupled mixed-layer (q-flux) ocean) 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 forcing.

Greenhouse gas forcing produces warming in the troposphere, cooling in the stratosphere, and 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 9.2.1c). The spatial pattern of 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). In contrast, sulphate aerosol forcing results in cooling throughout most of the atmosphere, with greater cooling in the NH due to its higher aerosol loading (Figure 9.2.1e, see Chapter 2), thereby partially offsetting the greater NH greenhouse gas induced warming. The combined effect of tropospheric and stratospheric ozone forcing (Figure 9.2.1d) is 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.

The responses to anthropogenic forcing described above are distinct from those due to the natural forcings. Solar forcing results in general warming of the atmosphere (Figure 9.2.1a) with a pattern of surface warming that is similar to that of greenhouse gas warming, but in contrast to the response to greenhouse warming, solar forced warming extends throughout the atmosphere. Volcanic forcing leads to a surface and tropospheric cooling and a stratospheric warming that peaks several months following a volcanic eruption (note that peak forcing by aerosols ejected into the stratosphere occurs several months after the eruption) and lasts for several years. Volcanic forcing also likely leads to a response in the atmospheric circulation in boreal winter (discussed below). Changes occur over the 20th century in this forcing because of the change in the frequency and intensity of volcanic eruptions over the century, resulting in greater volcanic forcing towards the end of the 20th century than early in the 20th century. This increase results in a small warming in the lower stratosphere and near the surface at high latitudes, with cooling elsewhere (Figure 9.2.1b). 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 9.2.1f). Results obtained with the CSIRO model (Figure 9.2.2) suggest that black carbon, organic matter and biomass aerosols would slightly enhance the Northern Hemisphere warming that is shown in Figure 9.2.1f. 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.

[INSERT FIGURE 9.2.1 HERE]

[INSERT FIGURE 9.2.2 HERE]

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 would 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 recently reversed since about 1990 (Pinker et al., 2005; Wielicki et al., 2005; Wild et al., 2005) (see also Chapter 3, Section 3.4.3). The consistency between the IPCC AR4 models and the more recent analyses is shown in Figure 9.2.3 where the top-of-atmosphere (TOA) outgoing shortwave flux from the models is compared with that measured by the ERBS satellite (Wong et al., 2005) and inferred from ISCCP FD data (Zhang et al., 2004). 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 most (but not all) of the AR4 models. The ISCCP FD flux anomaly is larger than that for ERBS reflecting the possible aliasing of the stratospheric aerosol signal into the ISCCP cloud properties. Overall, the downward trends inferred from ISCCP FD and the ERBS data agree, within their stated accuracies and are in even better agreement if one only considers tropical latitudes (Wong et al., 2005). 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 Chapter 3. (Chapter 3, Section 3.4.3). Nevertheless, the model-predicted trends and the trends based on observations encompass each other closely though they are slightly smaller in most models than in the ERBS observations (which are considered more accurate than the ISCCP FD data). Differences between simulated and satellite trends are significant for only a few models. This suggests that the temporal pattern of the effects of forcing (aerosols and land surface changes if included) and response (clouds and water vapor) on the outgoing shortwave flux is reasonable in most of the models.

[INSERT FIGURE 9.2.3 HERE]

Figures 9.2.1 and 9.2.2 also show that the spatial signature of a climate model's response is seldom very similar to that of the forcing. This comes about because climate system feedbacks vary spatially and because numerous climate processes (such as global circulation and convection) 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. 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 given climate model to a given forcing depends on its representation of these feedbacks and processes. 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; Stenchikov et al., 2006), greenhouse gas changes (e.g., Fyfe et al., 1999; Shindell et al., 1999; Rauthe et al., 2004), and possibly changes in solar irradiance (Shindell et al., 2001a; Tourpali, 2003; Egorova, 2004; Stendel et al., 2006) can alter the North Atlantic Oscillation or the Northern Annular mode. For example, volcanic 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. 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 temporal evolution of the different forcings (see Chapter 2, Figure 2.26) generally helps to distinguish the responses to the given forcings. For example, Santer et al. (1996b; 1996c) pointed 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 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 are optimally consistent with the observations. The fingerprints are typically estimated from ensembles of climate model simulations forced with reconstructions of past forcing. However, few studies have examined how uncertainties in the spatial pattern of forcing 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, 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 0.33 to 0.74 in different studies (Rotstayn and Penner, 2001; Chuang et al., 2002; Menon et al., 2002a; Rotstayn and Liu, 2003). The ratio of the ocean and land forcing for the sum of the first and second indirect effects ranges from 0.4 to 1.8 in different studies (Kristjansson, 2002; Lohmann and Lesins, 2002; Menon et al., 2002a). It may be possible to exclude some of these forcing patterns using spatially explicit satellite data, for example. But since it is not yet clear the contribution of these spatial uncertainties in the pattern of the forcing to the uncertainty of the response patterns needs to be considered.

9.2.2.3 Uncertainty in the temporal pattern of response

Climate model studies have generally not systematically explored the effect of uncertainties in the time-evolution of each of the forcings used in those studies. 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 largest uncertainties may be related to the anthropogenic emissions of 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. Fortunately, the time history of SO_2 emissions is better known. Overall, while the increasing forcing by greenhouse gases is relatively well known, the time evolution of forcing by other anthropogenic factors is more uncertain. However, the global mean forcing by many of these factors may have decreased or shown little change since the turn of the 21st century (see also Figure 9.2.3).

There are also very large uncertainties in the temporal forcing associated with solar radiation changes, particularly on timescales longer than the 11-year cycle. Previous estimates have used sun spot numbers to determine these slow changes in solar irradiance over the last few centuries, but are not necessarily supported by current understanding (Lean et al., 2002; Foukal et al., 2004; Chapter 6, Section6.4.1; 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 to a climate model and its corresponding responses. Nevertheless, any given model simulation will normally use forcings that have been developed from forward estimates.

It has been shown that for most forcings, the global average surface temperature response per unit forcing (or the "efficacy", see Chapter 2, Section 2.8.5) is similar within a given climate model (to within approximately 25% for most forcings) (Hansen et al., 1997; Forster et al., 2000; Rotstayn and Penner, 2001; Gregory et al., 2004a, Chapter 2, Section 2.8.5). 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. If the space-time pattern of the model response is incorrect, then the scaling will not necessarily encompass a value of one. Attribution studies evaluate the consistency between the model-simulated amplitude of response and that which is inferred from observations. However, for a model simulation to be considered consistent with the observations, the forcing used in the model should remain consistent with the uncertainty bounds from forward model estimates of forcing (see Chapter 2). In the case of uncertain forcings, scaling factors that are significantly different from one, on the other hand, should nonetheless provide information about the presence of the response pattern in observations that accounts for uncertainty in the strength of the 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. Variations in the time history 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. 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.

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. An explicit accounting of how uncertainties in the net forcing as a function of time contribute to uncertainties in the temporal evolution of temperature is not available. However, since models often employ different implementations of external forcing, the use of such different simulations for detection and attribution suggests that results are not very sensitive to small forcing differences. 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 versus the other as in the recent study by 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. However, detection results suggest that although the simulated responses to aerosol forcing are quite model dependent, this does not strongly affect estimates of the amplitude of the response to greenhouse gases (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. Forward model estimates of 20th century radiative forcing suggest that it is very likely that the total change in forcing is positive but less than 2.58 W/m^2. 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 is negative with magnitude very likely less than 1.7 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) and observations indicate a likelihood that aerosol forcings could indeed be less negative than -1.7 W/m^2 (see Table 2.12). The large uncertainty in total aerosol forcing inhibits attempts to accurately infer the climate sensitivity from observations (see Section 9.6). It also increases uncertainties in results that attribute cause to observed climate change (see Section 9.4.1.4), and is in part responsible for differences in projections of future climate change (see 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 relative uncertainties. Nonetheless, the different spatial patterns of forcing and response from these agents may offer opportunities to more precisely identify human influences on the climate system provided that it will be possible to determine their spatial/temporal pattern of emissions and response.

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 have been recognized as 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 industrial period, while the magnitude of that forcing is affected by uncertainty.

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. One example concerns difference in the space-time patterns of the response to greenhouse gas forcing and the response to sulphate aerosol forcing.. In contrast, the climate response and time evolution of other anthropogenic forcings is more uncertain, making simulated climate response and detection more difficult.The time-evolution, and to some extent the spatial pattern, of climate response to natural forcings is quite different from that of anthropogenic forcing. This makes it possible, with uncertainties, 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 do we Consider Pre-Industrial Climate Change?

The Earth system has undergone large-scale climate changes in the past (see Chapter 6) that hold important lessons for the understanding of present and future climate change. These changes resulted from the response of the climate system to natural external forcings, which sometimes trigger 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 too short to fully understand climate variability on interdecadal and longer timescales and is affected by anthropogenic as well as natural external forcings.

Changes in the more distant past, such as the LGM, provide opportunities to study the effects of major feedbacks in the climate system. Indirect indicators ("proxy data") must be used to infer past climate variations (see Chapter 6) prior to the instrumental era (see Chapter 3). These indicators include tree ring width and density, fossil corals, oxygen isotope ratios in ice cores and marine cores, pollen counts in sediment layers and sulphate deposition on ice caps. A complete description of these data and of their uncertainties can be found in Chapter 6.

Here we restrict the discussion mainly to the observations and simulations of the last millennium, so as to place 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. Increased ability to understand and simulate the climatic responses to these forcings in the past increases our confidence in our overall understanding of the impact of external forcing, including anthropogenic forcing, on the climate. In addition, the residual variability not explained by forcing may help to validate climate model internal variability. We also consider two time periods analyzed in the paleoclimate 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 show relatively strong changes compared to the present, and there is relatively good information from data synthesis and model simulation experiments (Braconnot et al., 2004; Cane et al., 2005).

Many simulations with models of intermediate complexity (EMICS) and comprehensive AOGCMs are available for all three of these periods, including an increased number of coupled simulations that have recently become available, run with the same model version as used for simulation of the climates of the 20th and 21st centuries. These simulations, and their comparison with proxy data, have improved our understanding of the role of the ocean and vegetation feedback in the response of the climate system to external forcing. They also provide new information on changes in short-term climate variability and climate teleconnections.

9.3.2 What can we Learn from the Last Glacial Maximum and the Mid-Holocene?

Relatively high quality global terrestrial 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 climate was much colder and drier during the LGM as is indicated by the extensive tundra and steppe vegetation that existed during this period. The CLIMAP (1981) reconstruction of LGM sea surface temperature has also been improved (see Chapter 6). 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 generally simulate 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 the PMIP results from simpler models that were discussed in the TAR (IPCC, 2001). Only one simulation exhibits a very strong response with a cooling of –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 observations, and better agreement is obtained with new coupled simulations using more recent models and, more complete feedbacks (see Chapter 6).

Closer to the present, during the mid-Holocene, one of the most noticeable changes in climate 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 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 to 10% larger than for atmosphere only simulations and are in better agreement with data reconstructions (Braconnot et al., 2004). Results for the Indian and Southwest Asian monsoon are less consistent, as is also the case projections of the future (see Chapter 10).

As noted in the TAR, vegetation change during the mid-Holocene is an important feedback in explaining the wet conditions that prevailed in the Sahel region (IPCC, 2001), as are ocean feedbacks (Ganopolski et al., 1998; Braconnot et al., 1999), although soil moisture may counteract some of these feedbacks (Levis et al., 2004). Wohlfahrt et al. (2004) showed that in mid- and high-latitudes the vegetation and ocean feedbacks enhance 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 then amplified when vegetation feedbacks are included (Wohlfahrt et al., 2005). This point needs to be considered when analysing future climate changes in these regions from model scenarios.

In addition, a wide range of proxies containing information about ENSO variability during the mid-Holocene is now available (see Chapter 6). These data suggest that ENSO variability was weaker prior to approximately 5,000 years before present than it is today (Moy et al., 2002 and references therein; Tudhope and Collins, 2003). Several studies have now 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 (e.g., Liu et al., 2000), for which simulations produce reduced variability in precipitation over most ocean regions in the tropics (Braconnot et al., 2004; Zhao et al., 2005). They also show a tendency for less frequent and intense ENSO events, in qualitative agreement with data, although there are large differences in magnitude and inconsistent responses of the associated teleconnections (Otto-Bliesner, 1999; Liu et al., 2000; Kitoh and Murakami, 2002; Otto-Bliesner et al., 2003). A key element of the ENSO response is the Bjerknes (1969) feedback mechanism. The increased mid-Holocene solar heating in boreal summer leads to more warming in the western than eastern Pacific, which strengthens the trade wind and inhibits the development of ENSO (Clement et al., 2000; Clement et al., 2004). One study also suggests that a change in the mean sea level pressure in the north Atlantic occurred corresponding to a negative NAO phase. This is consistent with data, although it is difficult to determine whether the more negative NAO phase suggested from data corresponds to a change in the mean state or to changes in decadal variability (Gladstone et al., 2005).

9.3.3 What can we Learn from the Past 1000 Years?

External forcing changes relative to the present are generally small for the last millennium when compared to those from 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 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 6.10, and the reviews by Jones et al., 2001 and Jones and Mann, 2004). Northern Hemisphere mean temperatures early in last the millennium were similar to or slightly cooler than those of first half of the 20th century. Temperatures decreased subsequently, and then rose rapidly during the most recent hundred years. This long term tendency is punctuated by substantial shorter term variability (see Chapter 6, Figure 6.10). The late 17th century, early 18th century and the period around 1830-1850 are the coldest of the last millennium, with temperatures 0.5–1°C below the 20th century mean value. New reconstructions suggest larger variations than contained in the Mann et al. (1999)

reconstruction (see Chapter 6), but uncertainty remains in the magnitude of inter-decadal to inter-century scale variability. This uncertainty arises from different sources depending on the proxy data and reconstruction methods used (see Chapter 6 for a complete discussion). Nonetheless, all reconstructions indicate that the Northern Hemisphere was likely warmer during the late 20th century than at any other time during the last millennium (see Chapter 6, Figure 6.10).

Simulations of the last millennium using reconstructed solar and volcanic forcings (Chapter 6, Figure 6.13 have been performed using a range of models, including some runs 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). Despite the differences between the various experiments, which result from the use of different models and the use of different natural forcing reconstructions (see Section 9.2), the evolution of the Northern Hemisphere annual mean surface temperature simulated by these models displays some common characteristics that are consistent with the broad features shown by data (Figure 9.3.1). In particular, all simulations suggest that the period around 1675 to 1715 was one of the coldest periods of the millennium, which is in qualitative agreement with the proxy reconstructions, and that the 20th century was warmer than the last 800 years for example, Stendel et al. (2006). The simulations also show that it is not possible to simulate the large warming during the 20th century without anthropogenic forcing (Crowley, 2000; Bertrand et al., 2002; Bauer et al., 2003; Hegerl et al., 2003), stressing the impact of human activity on the recent warming.

However, while there is broad qualitative agreement, there are differences among and between both simulations and proxy reconstructions of last millennium, particularly in the magnitude of past climate variations. The uncertainty in the magnitude of historical variations in the reconstructions makes it difficult to fully assess model simulated variability. The simulated magnitude of climate variations varies between models due to substantial differences in forcing and model formulation (for example, the sensitivity to external forcing varies by a factor of two between different climate models, see Chapter 8 , Table 8.8.1). 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. 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), and cannot account for differences between the simulations of Northern Hemisphere mean temperature. Other sources of uncertainty include the model ocean initial conditions which could explain some of 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., 2005). Despite uncertainties, residual variability that remains in high variance proxy reconstructions after estimates of the responses to external forcing have been removed provides a useful check on AOGCM simulated internal variability, and is broadly consistent with model simulated internal variability (e.g., Hegerl et al., 2003).

9.3.3.2 Role of volcanism and solar irradiance

Volcanic eruptions that eject large quantities of aerosol into the stratosphere cause rapid decreases in hemispheric and global mean temperatures followed by gradual recovery over several years. In addition, changes in the frequency of large eruptions result in climate variability on decadal to centennial time-scales (Crowley, 2000; Briffa et al., 2001; Bertrand et al., 2002; Bauer et al., 2003; Table 9.3.1). Hegerl et al. (2003), using a multi-regression approach (see Appendix 9.A.1, and 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 Northern Hemisphere mean annual mean temperature (Figure 9.3.1; Table 9.3.1) with high significance. Modelling studies also 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, although a reduction in solar forcing added 40% to the overall cooling.

[INSERT FIGURE 9.3.1 HERE]

[INSERT TABLE 9.3.1 HERE]

There is substantially 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, uncertainty also arises because the spatial response of surface temperature to solar forcing resembles that due to anthropogenic forcing (Nesme-Ribes et al., 1993; Cubasch et al., 1997; Rind et al., 2004; Zorita et al., 2005). Analyses that make use of differences in the temporal evolution of solar and volcanic forcing are better able to distinguish between the two (see also Section 9.4.1.4 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), although the effect of solar forcing has been detected over parts of the 20th century in some time-space analyses (see Section 9.4.1.4). Shindell et al. (2003) demonstrate in a model simulation that both volcanic and solar forcing are important globally, but that solar forcing plays a larger role in terms of regional anomalies because of dynamical feedbacks that have large regional projections. These uncertainties in interpretation of the role of different forcings reflects substantial uncertainties in our knowledge about the size of past volcanic forcing and of the timing and size of long-term variations in solar forcing, as well as differences in the way these effects are taken into account in model simulations.

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 ENSO variability do not appear to be correlated in an obvious way with mean state changes in the tropical Pacific or global mean climate. However, an empirical analysis of proxy-based reconstructions over the past three centuries (Adams et al., 2003) finds statistical evidence for an El Niño-like anomaly during the first few years following explosive tropical volcanic eruptions. A study with a simplified model of the tropical Pacific coupled ocean-atmosphere system supports the possibility of a link with volcanic forcing over the past millennium (Mann et al., 2005). However, substantial uncertainties remain as to whether and how ENSO changes in response to volcanism.

Extratropical variability also appears to respond to volcanic forcing. In particular, during the winter following a large volcanic eruption, the zonal circulation may be more intense (NAM/NAO-like response; Shindell et al., 2004; Yoshimori et al., 2005). This implies a relative warming over the continents during the cold season that could partly offset the direct cooling due to the volcanic aerosols (see Chapter 8, Section 8.4.1; Robock, 2000; Shindell et al., 2003). Some simulations also indicate a tendency towards the negative NAO state during periods of reduced solar input (Rind et al., 2004; Ruzmaikin et al., 2004; Stendel et al., 2006), as do reconstructions of this pattern for the northern hemisphere (Luterbacher et al., 2004). Note, however, that not all models show such a response (Palmer et al., 2004). This would imply a solar forcing role in some long-term regional changes and thus possibly a contribution to cooling over the Northern Hemispheric continent during some periods such as the cooling 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). The largest anthropogenic land cover changes involve deforestation (see Chapter 2), and the greatest proportion of deforestation has occurred in the temperate regions of the Northern Hemisphere (Ramankutty and Foley, 1999; Goldewijk, 2001). Some studies suggest that the effect on global mean radiative forcing of land cover change is comparable to that due to aerosols, ozone, solar variability and minor greenhouse gases (see Chapter 2; Section 9.2.1.1). Model simulations (e.g., Betts, 2000; Matthews et al., 2004) suggest that land use change may have had a cooling influence on climate, leading to a cooling of 1–2°C in winter and spring over the major agricultural regions of North America and Eurasia.

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 period in the Southern Ocean around the 14–15th centuries recorded in some proxies (Goosse et al., 2004), which would imply that the Southern Ocean has only partly responded to present-day changes in radiative forcing and therefore, that a large committed warming may remain to be realized (Weaver et al., 2000; Goosse and Renssen, 2001).

9.3.4 Summary

Considerable progress has been made since the TAR (IPCC, 2001) to better understand the response of the climate system to external forcings. Periods like the mid-Holocene and the Last Glacial Maximum are now seen as benchmarks for climate models that are used to simulate future climate (see Chapter 6). While considerable uncertainties remain in the climate reconstructions for these periods, and in the boundary conditions used to force climate models when simulating time slices during these periods, comparisons between simulated and reconstructed conditions in the LGM and Mid Holocene demonstrate that models capture the broad features of these climates. Modelling studies of these periods have also increased understanding of the roles of ocean and vegetation feedback in determining the response to solar and greenhouse gases forcing. Moreover, although proxy data on paleo-climate interannual to multidecadal variability remains 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 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 late 20th century temperatures were warmer than they have been for the last 1000 years (see Chapter 6). While uncertainties remain in the temperature and forcing reconstructions, and in the models used to estimate the responses to external forcings, the available detection studies, modelling and other evidence supports the conclusion that volcanic, and possibly solar forcing, has very likely affected Northern Hemisphere mean temperature variations over the past millennium and that external influences xplain a substantial fraction of interdecadal variability. AOGCMs, when driven with estimates of external forcing for the last millennium, simulate changes in hemispheric mean temperature that are largely in agreement with proxy reconstructions (given their uncertainties). Climate response to greenhouse gas increases can be detected in a range of proxy reconstructions by the end of the records.

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 (see 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 0.75°C by 2004 relative to 1860-1900 (calculated by fitting a low pass filtered time series, Chapter 3; and approximately 0.6°C per century over the 1901–2004 period if approximated by a linear trend, which under-estimates the observed change at the end of the record).Warming rates were greater after the mid 1970s with a warming rate of 0.15 to 0.18°C per decade over the 1979–2004 period (see Chapter 3). A 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 2000 years (see 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 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 (see Chapter 3, Figure 3.2.6). Since the mid-1970s, land regions have warmed at a faster rate than oceans in both hemispheres (see Chapter 3, Figure 3.2.8) and warming over the southern hemisphere was smaller than that over the northern hemisphere during this period (see Chapter 3, Figure 3.2.6), 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 coupled model simulations 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. Figure 9.4.1 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 model physics. Similarly, the effects of tropospheric and stratospheric ozone changes are included in some simulations but not others, whilst the naturally forced simulations include a variety of different representations of changing solar and volcanic forcing. Despite the inclusion of this additional uncertainty there is a clear separation in Figure 9.4.1 between the simulations with anthropogenic forcings and the simulations without. Larger inter-decadal 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-decadal variability that is simulated by models. Figure 9.4.1 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 9.4.1 HERE]

Figure 9.4.2 compares observed near –surface temperature trends over the globe (top panels) with those simulated by climate model simulations that include anthropogenic and natural forcing (middle panels) and the same trends simulated by climate models that include only natural forcings (lower panels). The observed trend over the entire 20th century (Figure 9.4.2 top left panel and Chapter 3, Figure 3.2.9) shows near-uniform warming with the exception of the Southeast United States, northern North Atlantic, and isolated gridboxes in Africa and South America; see also Chapter 3, Figure 3.2.9). 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 9.4.2, middle panels) 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 9.4.2, lower panels).

[INSERT FIGURE 9.4.2 HERE]

Global mean and hemispheric scale temperatures on multi-decadal time scales are largely controlled by external forcings. Stott et al. (2000) analysed an ensemble of integrations of HadCM3 (see Chapter 8, Table 8.2.1) including both anthropogenic and natural forcing that successfully simulates 20th century global mean and large scale land temperature variations. Calculations of the percentage of total variance explained by the model's response to external forcings indicate that the climate response on large spatial scales, particularly over land, is strongly influenced by external factors. This external control is demonstrated by ensembles of model simulations with identical forcings (whether anthropogenic or natural) whose members have 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).

Model simulations are consistent in showing that the global mean warming observed since 1970 can only be captured when models are forced with combinations of external forcings that include anthropogenic forcings (Figure 9.4.1). 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., 2005). In all cases the response to forcing from well-mixed greenhouse gases dominates the anthropogenic warming in the model. When the same models include natural forcings only, the observed warming is not reproduced. Therefore, modelling studies indicate that late 20th century warming is much more likely to be anthropogenic than natural in origin.

Modelling studies indicate more uncertainty over 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., 2003a; Nozawa et al., 2005) with some studies finding a greater role for solar forcing than other forcings before 1950 (Meehl et al., 2004) whilst others find that volcanic forcing (Broccoli et al., 2003; Hegerl et al., 2003) or natural internal variability (Delworth and Knutson, 2000) could be more important. There could also be an early expression of greenhouse warming in the early 20th century (Tett et al., 2002; Hegerl et al., 2003). Differences between simulations including increases in greenhouse gases only and runs also including the cooling effects of sulphate aerosols (e.g., Tett et al., 2002) indicate that the cooling effects of sulphate aerosols could 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. This model simulates regions of cooling in the mid century that are also observed and that are caused in the model by negative surface forcing in these regions from organic carbon and black carbon, mainly associated with biomass burning. Variations in the Atlantic Multidecadal Oscillation (see Chapter 3, 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) and 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.

The fact that climate models are only able to reproduce observed 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. Nonetheless, given the large uncertainties in aerosol forcings, the close level of agreement between many models and observations could have been obtained fortuitously 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). Multi-signal optimal 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, but determining their amplitudes from observations. Section 9.4.1.4 assesses studies of the relative contributions of the different forcing factors to global temperature changes.

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 the PCM model by Meehl et al. (2004) who demonstrated that the near-surface global mean temperature response to a combination of forcings is equivalent to the sum of the responses to the individual forcings. Gillett et al. (2004c) similarly found that the global mean and large-scale near-surface temperature responses to greenhouse gases and sulphate aerosols combine linearly in the HadCM2 model. Linear additivity was also 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 and al., 2003; Santer et al., 2003b). Meehl et al. (2003) found that additivitiy does not hold so well for regional responses in the PCM; the response to solar forcing in the PCM was enhanced regionally in the presence of greenhouse forcing, in part because altered cloud patterns affected the heterogeneity of the solar surface heating and altered regional feedbacks that depended on the climate base state in their model. Using an atmosphere only model forced by observed SSTs, Sexton et al. (2003) also found that the greenhouse gas and direct sulphate aerosol effect responses add linearly, although they also found 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..

9.4.1.3 Variability of temperature from observations and models

Year to year variability of global mean temperatures of the most recent models compares reasonably well with that of observations, as can be seen by comparing observed and modelled variations in Figure 9.4.1a. A more quantitative evaluation of modelled internal variability can be carried out by comparing the power spectra of observed and modelled global mean temperatures. Figure 9.4.3 directly 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 the simulations made as part of the IPCC AR4 20C3M project. Figure 9.4.3 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 9.4.4 shows that this is also the case regionally (see also Karoly and Wu, 2005), although in this case models generally simulate more variance than is observed on short timescales.

[INSERT FIGURE 9.4.3 HERE]

[INSERT FIGURE 9.4.4 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 (see Section 9.3.4). Chapter 8 discusses the simulation of major modes of variability in models, and the extent to which models can simulate them (including on decadal to inter-decadal time scales).

9.4.1.4 Detection, attribution, and quantification of the influence of external forcing on global surface temperature.

There are now a greater number of attribution studies than were available for the TAR, and these have used more recent climate data than previous studies and a greater variety of model simulations. Whereas many detection studies reported in the TAR considered data only until 1995 or 1996, most detection studies since the TAR have analysed data at least until the end of the 20th century with some analyses considering data from the early 21st century. The greater variety of model simulations includes more sophisticated treatments of a greater number of forcings of both anthropogenic and natural origins.

The longer record, improved models, and strengthening anthropogenic signal has increased confidence in detection of an anthropogenic signal in the instrumental record (see, for example, the recent review by IDAG, 2005). Optimal fingerprinting 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 (Hegerl et al., 2001; Gillett et al., 2002b; Tett et al., 2002; Zwiers and Zhang, 2003; Allen et al., 2005; IDAG, 2005; Stott et al., 2006b; Stott et al., 2006c; Zhang et al., 2006). An analysis using an EBM to infer the likely observed global mean responses to natural and anthropogenic forcings (Stone and Allen, 2005a) support this conclusion. Recent statistical analyses of the observational record also increase confidence. For example, Fomby and Vogelsang (2002), using a serial-correlation robust test of trend, find 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. Kaufmann and Stern (2002) detected a significant human influence on near-surface temperatures by analysing the lagged covariance structure of hemispheric mean temperatures and 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).

Since the TAR there has 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 factors such as from volcanic eruptions and changes in solar output. An example is Tett et al. (2002), who simulated the climatic response to natural and anthropogenic forcings from 1860 to 1997 using the HadCM3 model. By analysing observed and modelled near-surface temperature changes using an optimal detection methodology, they detected the effects of changes in well-mixed greenhouse gases, other anthropogenic forcings (mainly the effects of sulphate aerosols on cloud albedo) and natural forcings, showing that all have had a significant impact on 20th century temperature changes. They estimated the linear trend in global-mean near-surface temperature from well mixed greenhouse gases to be 0.9 ± 0.24°C per century, offset by cooling from other anthropogenic forcings of 0.4 ± 0.26°C per century giving a net anthropogenic warming trend of 0.5 ± 0.15°C per century. Over the entire century natural forcings gave a linear trend close to zero, with a cooling trend over the latter part of the century following an earlier warming trend. Their analysis suggests that the early 20th century warming can best be explained by a combination of warming due to increases in greenhouse gases and natural forcing, some cooling due to other anthropogenic forcings, plus a substantial, but not implausible, contribution from internal variability. In the second half of the century they found that the warming was largely caused by changes in greenhouse gases, with changes in sulphates and, perhaps, volcanic aerosol cooling offsetting approximately one-third of the warming.

This analysis, in common with most analyses up to that point, did not take account of sampling uncertainty in the modelled signals of climate change. This uncertainty, which results from the influence of internal variability on signal estimates from finite-member ensembles, leads to a low bias in estimates of the scaling factors by which the modelled response to a particular forcing must be scaled up or down to best match the observed change (see Appendix 9.A.1). This bias is likely to be greater for weak forcings (Allen and Stott, 2003; Stott et al., 2003b). On the whole, taking account of sampling uncertainty (as developed by Allen and Stott, 2003, and implemented by Stott et al., 2006c and many other studies) 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.

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 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 between –0.1 and 0.1°C (results labelled F(22), F(21), and F(19) Figure 9.4.5 middle panel). Scaling factors for the three forcings are shown in Figure 9.4.5 (top panel). A similar analysis for the MIROC model finds a slightly larger warming contribution from greenhouse gases of between 1.2 and 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 (results labelled F(A2) in Figure 9.4.5). 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 9.4.5, bottom panels). These results, obtained from four different climate models, with different sensitivities and forcings, are largely consistent with those obtained by Tett et al. (2002) for the HadCM3 model, indicating that observational constraints reduce the uncertainty on warming rates attributable to greenhouse gas increases.

[INSERT FIGURE 9.4.5 HERE]

Not all models have the full range of simulations required to directly estimate the responses to individual forcings that are required for multi-signal detection and attribution anayses. 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 the observed global mean time series using optimal detection. By tuning an EBM to the observations, thereby using an AOGCM model solely to estimate internal variability, Stone and Allen (2005a) 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. Whereas this approach, based on an EBM, analyses changes in global mean temperature and does not consider spatial patterns of temperature change, a possible technique for estimating the spatio-temporal patterns of response to each forcing is to regress the time information of temperature at each spatial point against the timeseries of the individual forcings (Crooks and Gray, 2005). Fingerprints for each forcing could then be derived for the full range of models in the IPCC AR4 20C3M project and compared with results from the full space-time analyses shown in Figure 9.4.5.

Gillett et al. (2002b) combined results from five models in a single analysis. They calculated the mean response patterns from the five models (HadCM2, HadCM3, CGCM1, CGCM2 and ECHAM3) which they used as fingerprints for a detection of greenhouse gas and sulphate aerosol influence, including an estimate of model uncertainty. Their results indicate that inter-model differences do not greatly increase detection and attribution uncertainties as applied to temperature data, and that averaging fingerprints improves detection results. Gillett et al. obtained their estimate of model uncertainty 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. Huntingford et al. (2006) developed a more sophisticated methodology for incorporating modelling uncertainty into detection analyses by including an estimate of the inter-model covariance structure in the regression method. Their results (labelled "EIV" in Figure 9.4.5) support those results generated by spatio-temporal analyses on single models (labelled "F" in Figure 9.4.5) in showing that it is likely that greenhouse gases alone would have caused more warming than has been observed over the 1950–1999 period.

Detection of a greenhouse gas influence on global mean temperature is a robust feature of a wide range of detection analyses. Optimal detection analyses available to the TAR, which quantified the contributions to past climate change from greenhouse gases and other external forcing factors, showed that it was likely that greenhouse gases were responsible for at least half of the warming observed over the last 50 years of the 20th century. The analyses performed since the TAR 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 the 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 (Santer et al., 1996b; Santer et al., 1996c; Stott et al., 2006c), in addition to the trend in the hemispheric warming contrast itself, whose strength is model dependent (Hegerl et al., 2001). 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, 2005; Nagashima et al., 2006).

9.4.1.5 Bayesian detection and attribution studies.

All studies described in Section 9.4.1.4 employ the standard frequentist approach to assessing hypotheses for causes of past climate change (see Appendix 9.A.2). However, there is also a developing interest in applying Bayesian methods which allow the inclusion in the analysis of prior information. In a Bayesian framework all the above studies assume a non-informative prior distribution for the scaling factors, equivalent to saying that we have no previous knowledge of the values of the scaling factors. Some recent studies that have used informative prior distributions have attempted to take prior and other sources of uncertainty into account in a Bayesian framework. Most studies use Bayes Factors (ratios of posterior to prior odds) to assess evidence supporting competing hypotheses. Calibrated descriptors, such as "decisive", "very strong", "strong" or "positive", are used to describe different ranges of Bayes Factors (Kass and Raftery, 1995, describes Bayes Factors and the descriptors in detail).

Schnur and Hasselman (2005) analysed recent 31-year spatial trend patterns (c.f. Hegerl et al., 1997) of patterns of near-surface temperature in a Bayesian framework in order to distinguish between three competing hypotheses, namely that the climate changes observed late in the 20th century can be explained by internal variability alone, by internal variability and greenhouse gas forcing (G), or by internal variability and the combined effect of greenhouse gas and sulphate aerosol forcing (GS). By analysing simulations from a number of different climate models and assuming that these three possibilities were equally likely a priori, Schnur and Hasselman inferred that the odds of hypotheses G and GS against the internal variability hypothesis are very large. This is equivalent to saying that the G and GS signals are clearly detected in the observations. However, they also found that GS is only twice as likely as G, which does not represent decisive evidence for one hypothesis over the other. Whilst they postulate that lack of decisive evidence for a sulphate aerosol effect is due to taking account of modelling uncertainty in the response to aerosols, two other studies which include modelling uncertainty find a clear detection of sulphate aerosols, suggesting that the use of multiple models helps to reduce uncertainties and makes detection of a sulphate aerosol effect more likely (Gillett et al., 2002b; Huntingford et al., 2006).

Min et al. (2004) applied a Bayesian analysis to 1979–1999 surface and 70 hPa NH temperature from reanalysis and found evidence that simulations with anthropogenic forcings explain observations better than a control simulation. Min and Hense (2006), using the the IPCC AR4 20C3M simulations, find decisive evidence for an "all forcings" explanation over the 20th century, as against alternative explanations involving internal variability alone or in combination with either natural or greenhouse gas forcing. They also found strong evidence for both "all forcings" and naturally forced explanations of temperature changes during the 1900–1949 period.

Differences in separate detection of sulphate aerosol influences in a multi-signal approach can reflect differences in the diagnostics applied (e.g., the space-time analysis of Tett et al (1999) vs. the space only analysis of Hegerl et al. (1997; 2000) as was shown by Gillett et al., (2002a) (see also Hegerl and Allen, 2002). One Bayesian study of spatial patterns of trends finds that the evidence of a significant aerosol cooling effect on past temperature changes is relatively weak compared to the very strong evidence for an anthropogenic influence (from the combined effects of greenhouse gases and aerosols, Schnur and Hasselmann, 2005), although another Bayesian study of temporal change in hemispheric mean temperatures finds more decisive evidence for an aerosol cooling effect (Smith et al., 2003).

Lee et al. (2005) applied a variant of the Bayesian technique described by Berliner et al. (2000) to a conventional optimal fingerprinting in order to evaluate detection and attribution hypotheses about the GS scaling factor. Lee et al. (2005) evaluate the evidence for the presence of the GS signal, estimated from CGCM1 and CGCM2 (see Chapter 8, Table 8.1, McAvaney et al., 2001, for model information), in observations for several 5-decade windows, beginning with 1900–1949, and ending with 1950–1999. Evidence supporting detection was assessed by requiring a high posterior likelihood that the GS scaling factor was greater than 0.1. Evidence supporting attribution was similarly assessed by requiring a high posterior likelihood that the scaling factor was within 20% of unity. This study found very strong evidence in support of detection during both halves of the 20th century regardless of the choice of prior distribution. On the other hand, evidence for attribution, if as stringently defined as above, is weak. Positive evidence for attribution was obtained when using noncommittal priors and a less stringent attribution criterion that requires the model response to the GS forcing to be within 50% of the apparent observed response. Lee et al. (2005) estimated that strong evidence for attribution may emerge within the next two decades as the anthropogenic signal strengthens.

In a further study using Bayesian techniques, Lee et al. (2006) have considered predictability of decadal mean temperature anomalies that arises from anthropogenic forcing. 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 hindcast global mean temperature anomalies (Figure 9.4.6).

[INSERT FIGURE 9.4.6 HERE]

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 (see Section 9.4.1.4) in turn determines observational constraints on the transient climate response and therefore on likely future rates of warming. Scaling factors derived from detection analyses can be used to scale predictions of future change by assuming that fractional error in model predictions of global mean temperature change is constant, in a manner analogous to the Model Output Statistics (MOS) technique (Klein and Glahn, 1974) which has a long history of application in weather forecasting (Allen et al., 2000; Allen et al., 2002; Allen and Stainforth, 2002; Stott and Kettleborough, 2002). In this approach based on optimal detection, which is compared with other approaches for producing probabilistic projections in Section 10.5.4.5, different scalings are applied to the response to greenhouse gases and other anthropogenic forcings (noteably aerosols). Uncertainties calculated this way are likely to be more reliable than ranges of forecast uncertainty based on "ensembles of opportunity" (Allen and Stainforth, 2002) composed of simulations from 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. Predictions from these models produced in this way were shown to be relatively insensitive to the choice of three models used to produce them. 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 to lie between 1.5°C and 2.8°C (5 and 95 percentiles, see Section 9.6.1, Figure 9.6.2). Such approaches have also been used to provide observationally constrained predictions of global mean (Stott and Kettleborough, 2002) 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 11.2.2.2.4.

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, and the temperature contrast between Northern and Southern hemispheres, Braganza et al. (2004) found 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 optimal detection techniques (see 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 during prior to 1980 (see 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 underestimated (Stone and Weaver, 2002, 2003; Braganza et al., 2004). This underestimate is associated with an overestimate by models of the observed increase in daily maximum temperature. Braganza et al. (2004) showed evidence that this overestimate could be associated with models underestimating observed increases in cloud cover (since the middle of the last century over many continental regions and over the last 30 years over the oceans, see Chapter 3, Section 3.4.3.1), a result supported by other analyses (Dai et al., 1999; Stone and Weaver, 2002, 2003).

9.4.1.8 Remaining uncertainties

A larger range of forcing combinations has been analysed in detection studies than were available for the TAR. Nevertheless studies have concentrated on what are believed to be the most important forcings with most analyses excluding some forcings that could potentially have significant effects, particularly on regional scales but possibly on global scales also. Observational campaigns have demonstrated the importance of black carbon in the South Asia region (e.g., Ramanathan, 2005) and modelling studies have shown that the global forcing from black carbon could be greater than 0.5 W m^{-2} (Jacobson, 2001; Hansen and Nazzarenko, 2004), yet few detection studies have explicitly included the temperature response to black carbon aerosols because there are few transient coupled model simulations including this forcing due to large modelling uncertainties (see Section 9.2, Roberts and Jones, 2004). Land use changes are another forcing that could be potentially important, particularly on regional scales, and could have caused a cooling in winter and spring in northern Europe due to deforestation in pre-industrial times (see Section 9.3). Forcing from surface albedo changes due to land use changes is expected to be negative globally (see Chapter 2, Section 2.5.3 and Section 9.2.1.1) although tropical deforestation could increase evaporation and warm climate (see Chapter 2, Section 2.5.5) counter-acting cooling from albedo change; however the error in attribution of global mean temperature changes due to neglecting land use changes is likely to be relatively small. Attribution analyses that use recent model simulations which include carbonaceous aerosols and land use changes, such as the MIROC model (model ID 19, Chapter 8, Table 8.2.1; results labelled “F(19)” in Figure 9.4.5) and the HadGEM1 model (Stott et al., 2006b), continue to detect a significant anthropogenic influence on 20th century temperature changes. Other forcings discussed in Chapter 2 but not yet included in attribution analyses include stratospheric water vapour changes.

A sensitivity study incorporating a near-surface temperature pattern of response to black carbon estimated from a climate model with a slab ocean (Jones et al., 2005) indicated that the patterns of near-surface temperature response to sulphate aerosol and black carbon were so similar on large spatial scales (but of opposite signs) that a detection analysis was unable to distinguish them (see also Section 9.2). This suggests that attribution of the net aerosol contribution to past global-mean temperature change should be fairly robust to the inclusion of black carbon in future analyses but that separating sulphate cooling from warming due to black carbon would require a careful choice of diagnostics. However there are likely to be greater differences in the patterns of response in some regions such as India (Ramanathan, 2005), China (Roberts and Jones, 2004) and over the Arctic from changes in snow and ice albedos (Hansen and Nazzarenko, 2004) as well as in the vertical structure of the temperature response (see Section 9.2.2).

For each forcing considered, there are uncertainties associated with the temporal and spatial nature of the forcing and the modelled response. Progress has been made since the TAR in relaxing some of the previous assumptions that have been taken, including sampling uncertainty associated with estimating the climate's forced response from a limited number of simulations (Allen and Stott, 2003). One recent study has developed a methodology to take account of errors in the patterns of response to different forcings (Huntingford et al., 2006). However estimation of this modelling uncertainty is dependent on the model simulations available and large multi-model ensembles (Murphy et al., 2004; Stainforth et al., 2005) using different models. There remains considerable uncertainty in the forcings that are used in climate models. Since the TAR, estimates of the uncertainties in reconstructions of past solar forcing have increased (see Chapter 2 and Gray et al., 2005), and chemical and dynamical processes associated with the stmosphere's response to solar irradiance are omitted or not adequately resolved in many climate models used in detection studies (Gray et al., 2005). Also, 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). Furthermore, some models include the indirect effects of 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).

Optimal detection analyses take account of systematic biases in the patterns of climate response (which could result from model errors due to poor representation of feedbacks or errors in the imposed external forcing) by the inclusion of scaling factors. 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. However, most analyses do not fully account for systematic biases since they do not account for errors in the patterns of response to different forcings, although some recent analyses are beginning to take such uncertainties more fully into account by sampling multi-model ensembles (Schnur and Hasselmann, 2005; Huntingford et al., 2006). Nevertheless, the level of consistency between attribution results derived from different models (as shown in Figure 9.4.5), taken together with the ability of climate models to simulate large-scale temperature changes during the 20th century (see Figures 9.4.1 and 9.4.2), indicate that such model errors are likely to have a relatively small impact on attribution results of large scale temperature change at the surface.

A further source of uncertainty derives from the estimates of internal variability (including ENSO) that are required for all detection analyses. These estimates are generally model-based because of difficulties in obtaining reliable natural variability estimates from the observational record on the space and time scales considered in detection studies. However, models would need to underestimate variability by large factors to nullify detection of greenhouse gases in near surface temperature data; even the standard deviation of HadCM3 variability, which has larger variability than many models (see, e.g., Figure 9.4.3), would need to be underestimated by a factor of over 2 (Tett et al., 2002). This, on the other hand, appears unlikely given the quality of agreement between models and observations on the global scale (Figure 9.4.3) 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 (an increase of 40% in the estimate of the standard deviation 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 (see 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 analyses of free atmosphere temperature changes and these are discussed in Section 9.4.4.

9.4.2 Regional Surface Temperature Change

9.4.2.1 Observed changes

In Central Europe, where there is a centuries-long instrumental record, late 20th century temperatures are the warmest observed, for at least 250 years (see Chapter 3, Box 3.5.5 and Figure 3.8.6). Over the 1901–2003 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 (see Chapter 3, Figure 3.2.8 and Section 3.2.2.7, see also Figure 9.4.2). Warming has been strongest over the continental interiors of Asia and Canada and some mid-latitude ocean regions of the southern hemisphere. Since 1979, all land areas show warming except for a small region at the tip of South America (see Chapter 3, Figure 3.2.9). 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 (Section 3.4.3.1).

9.4.2.2 Studies based on space time patterns

The sensitivity of detection results to the space and time scale considered was discussed in the TAR. Idealised studies (e.g., Stott and Tett, 1998) showed that surface temperature changes are detectable on large spatial scales of the order of several thousand kilometres. The large-scale coherence of surface temperature changes means that significant warming trends are also detectable in observations at a much larger fraction of individual climate model grid boxes than can be explained by chance, calculated by a field significance test (Karoly and Wu, 2005). Furthermore, models typically do not underestimate natural internal variability at grid box scale (Karoly and Wu, 2005) or regionally over land (Figure 9.4.4), although the processes that generate variability on small spatial scales may not be well represented in models. Also, detection at an individual grid box does not necessarily imply that temperatures are controlled largely by external forcings and are therefore predictable at that grid box, as can be the case for larger area averages (Stott et al., 2000).

Previous global-scale analyses using space-time detection techniques (see Section 9.4.1.4) have robustly identified the influence of anthropogenic forcing on the 20th century global climate. Given this success, a number of studies have now extended these analyses to 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.

An approach taken by IDAG (2005) was to compare analyses of full space-time fields with results obtained after removing temporal information, either by removing the linear trend in the global mean or by removing the annual global mean from each year in the analysis. By doing so, 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.

Another approach for assessing the regional influence of external forcing is to apply detection and attribution analyses to observations in specific regions. Zwiers and Zhang (2003) assess the detectability of the GS signal as estimated by the Canadian Centre for Climate Modelling and Analysis (CCCma) CGCMs 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, see Chapter 8, Table 8.1, IPCC, 2001, and HadCM3, model ID 22, Chapter 8, Table 8.2.1) and also some sub-continental regions. They find evidence that climates in both continental domains have been influenced by anthropogenic emissions during the latter half of the 20th century, and generally also in the sub-contental domains (Figure 9.4.7). This finding is robust to the exclusion of North Atlantic Oscillation (NAO/AO) related variability (Thompson and Wallace, 1998), which could itself be related to anthropogenic forcing (see Fyfe et al., 1999; Shindell et al., 1999; Gillett et al., 2000; Gillett et al., 2003b) (see 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 9.4.7) becomes larger, reducing the signal to noise ratio in detection and attribution results (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. Therefore, signal to noise ratios vary between regions. Most of the results noted above hold even if the standard deviation of the internal climate variability from the control simulation is doubled.

[INSERT FIGURE 9.4.7 HERE]

Stott (2003) analysed HadCM3 simulations to estimate the influence of natural and anthropogenic forcings on 20th century near-surface temperature in six continental scale regions, each composed of a small number of sub regions. The warming effects of increasing greenhouse gas concentrations are detected in all the regions examined. In most regions, cooling from sulphate aerosols counteracts some of the greenhouse warming. However, the separate detection of sulphate aerosol signal in regional analyses remains difficult because of weaker signal to noise ratios and greater modelling uncertainty (Zhang et al., 2006). HadCM3 reproduces many features of the observed temperature changes and variability in the different regions (IDAG, 2005) and 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., 2005). Other studies show success at simulating regional temperatures when models include anthropogenic and natural forcings. Wang et al. (2006) showed that all the IPCC AR4 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-century warming event.

A comparison between a large ensemble of simulations of regional temperature changes made with a range of different models (from the same simulations from which the global mean temperatures are plotted in Figure 9.4.1) shows that the spreads of the multi-model ensembles encompas the observed changes in regional temperature changes in almost all regions (Figure 9.4.8). 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. These results, taken together with the sub-global detection analysis described above, provide evidence of a likely human influence on regional climates.

[INSERT FIGURE 9.4.8 HERE]

Nevertheless, some difficulties remain in simulating temperature changes in some regions of the World, although observed regional trends and variations could be wholly or partly caused by natural internal variability, which a particular model simulation could not be expected to reproduce. Parts of North America are not particularly well simulated by either HadCM3 or GFDL CM2, and GFDL CM2 underestimates warming in Northern Asia (Knutson et al., 2005) while HadCM3 fails to capture features of the observed warming in mid Asia (IDAG, 2005; Figure 9.4.8). An analysis of the IPCC AR4 20C3M experiments indicates that there is a large spread in simulations of 20th century temperatures in the central United States, a region that experienced rapid warming between 1901 and 1940 and rapid cooling between 1940 and 1979, suggesting that multi-decadal internal variability could be responsible (Kunkel et al., 2006). Some regional differences could be related to missing forcings and processes (e.g., the effects of land use changes) which could be locally important (see Chapter 2). Ramanathan et al. (2005) found that observed temperature trends in South Asia and the Northern Indian Ocean can be simulated with the PCM if atmospheric brown clouds are included in the model.

Optimal detection analyses can now also be applied to smaller, sub continental areas. An analysis of France (Spagnoli et al., 2002)indicated the potential for detecting changes at the country level, for some specific indices; for France detection of anthropogenic influence was seen for 30 year trends of summer daily minimum temperatures but not for summer daily maximum temperatures or winter temperatures. Stott et al. (2004) detect an anthropogenic influence on European summer mean temperature changes of the past 50 years, and Gillett et al. (2004a) detect an anthropogenic contribution to summer season warming for regions

of Canada prone to forest fires. Zhang et al. (2006) demonstrate the detectability of the G and GS signals in annual mean, and some seasonal mean temperatures, in Europe, Canada and China using a multi-model approach. Further evidence for an anthropogenic influence on regional climate comes from Min et al. (2005), who analyzed East Asian temperature changes in a Bayesian framework and showed strong evidence for detection with high Bayes factors (see discussion in Section 9.4.1.5) in this region. Min and Hense (2006) similarly showed that over most of the regional areas they considered, a Bayesian decision approach claisified temperature changes as being anthropogenically, rather than naturally, caused.
An alternative approach is to estimate spatio-temporal patterns of near-surface land temperature response with atmosphere-only general circulation model simulations forced with observed sea surface temperatures. Sexton et al. (2003). The smaller variability of atmosphere-only models compared to coupled models enables the detection of weaker signals and potentially the detection of anthropogenic effects on smaller spatial and temporal scales. This approach can also account for possible non-linear interaction between forcings, although the causes of changes in SSTs remain unexplained using this methodology.

9.4.2.3 Studies based on indices for temperature change

Indices of North American continental scale temperature change were analysed by Karoly et al. (2003). Observed trends in the indices, which included the regional mean, the mean land-ocean temperature contrast, and the annual cycle, were found 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. Changes in Australian mean, daily maximum and daily minimum temperatures and diurnal temperature range were analysed by Karoly and Braganza (2005b) using 6 coupled climate models. They showed that it is likely that there has been a significant contribution to observed warming in Australia from increasing greenhouse gases and sulphate aerosols. Nicholls (2003) showed that there has been an anomalous warming over Australia over the last few decades associated with a changed relationship between annual mean maximum temperature and rainfall since the early 1970s. 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. A recent warming trend not associated with rainfall variations that could therefore be associated with the enhanced greenhouse effect, has also been identified in New South Wales (Nicholls et al., 2005). By removing the rainfall related component of Australian temperature variations, thereby enhancing the signal-to-noise ratio, Karoly and Braganza (2005a) 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. Douville (2006) showed that removing the rainfall-related component of temperature over the Sudan and Sahel region improved the agreement between model simulations and observations of temperature change in this region over the last 60 years and proposed that applying the same methodology over other regions where precipitation is likely to affect the surface energy budget (Trenberth and Shea, 2005) would improve the detectability of regional temperature signals.

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 gradual reduction in the number of frost days over most of the mid-latitudes in recent decades, an increase in the number of warm extremes, particularly warm nights, and less so, hot days; and a reduction in the number of cold extremes at the daily timescale. Heat waves have increased in frequency during the latter part of the 20th century. 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 an important component of changes in the number of frost days and warm nights. Tebaldi et al. (2006) found that eight CGMs run for the IPCC AR4 agreed well with the observations that there has already been a trend in temperature-related extremes in the positive direction for heat waves and warm nights, and in the negative direction for 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. The indices proposed by Hegerl et al. (2004) were the average of the 30, 10, 5, and 1 hottest and coldest days of the year and Hegerl et al. (2004) found subtle but significant difference between changes in these indices and seasonal means over large areas of the globe, particularly over Europe, with signal-to-noise ratios for changes in these indices nearly as large as for changes in mean temperature. Christidis et al. (2005) detected robust anthropogenic changes in indices of extremely warm nights, 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, although less convincingly, and with model simulations underestimating the observed changes, significantly so in the case of the coldest day of the year. Christidis et al. (2005) found observataions were inconsistent with model simulations of extremely warm days with no detection of a significant human influence, which had the smallest signal-to-noise ratios of the four types of indices they examined.

9.4.3.3 Assessment of regional temperature extremes

It is important to be clear about the difference between the actual observed change in the variable of interest and the expected underlying change (Allen, 2005). Most of the variability of global mean temperature on multi-decadal timescales is externally driven (Stott et al., 2000) with global mean temperatures varying from year to year around the underlying forced changes due to internal variability (see Section 9.4.1). However, for regional temperature changes and for temperature extremes, the effect of internal variability becomes more important relative to the potentially predictable changes from external forcings. Most attribution studies have considered underlying deterministic changes rather than the actual events themselves. However, many important impacts of climate change are likely to manifest themselves through a change in the frequency or likelihood of occurrence of events that, taken individually, could be explained as naturally occurring (e.g., Palmer, 1999). Palmer and Raissanen (2002) assess how future increases in greenhouse gas forcing may change such risks in the case of extreme seasonal precipitation.

Allen (2003) and Stone and Allen (2005b) proposed a methodology for making quantitative attribution statements about specific types of climatic events, by expressing the contribution of external forcing to the risk of an event exceeding the observed magnitude. These studies proposed that the concept of the fraction of attributable risk (FAR), an established concept in epidemiological studies, should be applied to the problem of attributing a change in risk of a specific type of event to external forcing. If P_1 is the probability of a climatic event (such as a flood or heat wave) occurring in the presence of anthropogenic forcing of the climate system, and P_0 is the probability of it occurring if such anthropogenic forcing had not been present, then the fraction of the current risk that is attributable to past greenhouse gas emissions is given by $1-P_0/P_1$ (Allen, 2003). Analyses of attributable risk provide the basis for such statements as “half the deaths due to X are attributable to environmental risk factor Y” and are subject to well-documented hazards of interpretation (Greenland and Robins, 1988) which need to be borne in mind as they are extended to the climate problem. The “frequency of occurrence” interpretation becomes more problematic when dealing with the most extreme events that, by definition, occur very infrequently in both present-day and pre-industrial climate where the probability of events is much more difficult to assess.

Stott et al. (2004) investigated to what extent climate change could be responsible for the high summer temperatures in Europe during the summer of 2003 (described in detail in Chapter 3, Box 3.5) by applying the FAR concept to mean summer temperatures of a large region of continental Europe and the Mediterranean. Luterbacher et al. (2004) showed that the summer of 2003 in central Europe was likely the warmest in at least 500 years. Schär et al. (2004) showed that the central European heat wave of 2003 might be consistent with model predicted increases in temperature variability due to soil moisture and vegetation feedbacks, and hence greater likelihood of extremes. Overlain on an overall warming trend affecting extreme summer temperatures could be an influence of basin-scale changes in the Atlantic Ocean, related to the Meridional Overturning Circulation (MOC), which could drive multi-decadal variations in western European summer climate (Sutton and Hodson, 2005). 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 and Scherrer (2005) found evidence for a weak increase in European temperature variability in summer (and a decrease in winter) for the period 1961-2004, although these changes were not found to be statistically significant. Meteorological aspects of the summer 2003 heat wave are discussed in Chapter 3, Box 3.6, and in more detail by Fink et al. (2004) and Black et al. (2004)

There could be a change in extreme events (with impact-related thresholds being exceeded) both as a result of changes in mean temperatures without any increase in variability, or as a result of increases in variability alone, or a combination of both. Contrary to analyses of smaller continental regions (Hegerl et al., 2004; Schär et al., 2004), Stott et al. (2004) found no evidence for a change of variability in the future under the much larger region and for the seasonal mean change they considered and were therefore able to consider the problem as a two stage problem. Note that whilst several studies have indicated increased variability in European summer temperatures, the magnitude of the increase is model dependent and a weaker tendency to increased variability has been found in the Mediterranean region (see Chapter 11, Section 11.3.3.3.2). By carrying out an optimal detection analysis and comparing observed and modelled European temperature changes, Stott et al (2004) estimated the changes in 1990s temperatures in their large European region that were attributable to both anthropogenic and natural factors, and compared this with the temperature change attributable to natural factors alone. They then further used the model to estimate the probability of exceeding a particular extreme threshold in the presence and absence of the model simulated anthropogenic climate change after carefully evaluating the model's ability to simulate temperature variability in the region of interest. The FAR was then calculated from the ratio of these two probabilities. In this way they estimate, using a threshold that was exceeded in 2003, but in no other year since 1851, that it is very likely (a better than 9 in 10 chance) that past human influence has more than doubled the risk of a regional scale heat wave of at least this magnitude. Figure 9.4.9a shows the estimated likelihood of the risk (probability) of exceedance of a 1.6°C threshold (relative to the 1961-1990 mean) in the presence (red curve) and absence (green curve) of anthropogenic change, expressed both as a frequency (number of occurrences per thousand years). The clear shift from the green to the red distribution (Figure 9.4.9a) implies an FAR distribution whose mean is 0.75, corresponding to an increase in risk of a factor 4 (Figure 9.4.9b). This study considered only continental mean seasonal mean temperatures and consideration of shorter term and smaller scale heatwaves will require higher computational resolution and will need to take complexities such as land surface processes into account (Schär and Jendritzky, 2004).

[INSERT FIGURE 9.4.9 HERE]

9.4.4 Free Atmosphere Temperature

9.4.4.1 Observed changes

Observed free atmosphere temperature changes are discussed in detail in Chapter 3, Section 3.4.1 and a comprehensive review has been made by CCSP (2006); they are summarised briefly here. Radiosonde based observations (since 1958) and satellite based temperature measurements (since 1979) 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 in the troposphere being slightly more than at the surface. Since, 1979, due to differences between tropospheric datasets, it is not clear whether the troposphere has warmed more than or less than the surface. Global mean tropospheric temperatures increased at about 0.14°C per decade since 1958, and between 0.10°C and 0.20°C per decade since 1979. In the tropics, temperature increased at about 0.13°C per decade since 1958 and between 0.02°C and 0.19°C per decade since 1979. Whilst all datasets show that the stratosphere has cooled considerably from 1958 and from 1979 to present, there are large differences in the linear trend values from different datasets. However, a linear trend is a poor fit to the data in the stratosphere and the tropics at all levels (see 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 CCSP (2006). Errors remain in homogenized radiosonde datasets which could result in a net spurious cooling in the tropical troposphere and differences between different versions of tropopsheric satellite data result primarily from differences in how data from different satellites are merged together.

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. Analyses of radiosonde data have documented increases in tropopause height over the past 3–4 decades (Highwood, 2000; Seidel et al., 2001). Similar increases have been inferred from three different reanalysis products (ERA-15, ERA-40 and NCAR-NCEP) (Kalnay and al., 1996; Gibson and al., 1997; Simmons and Gibson, 2000; Kistler and al., 2001) and from model simulations with combined anthropogenic and natural forcing (Santer et al., 2003b; Santer et al., 2003c; Santer et al., 2004; see Figure 9.4.10). 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., 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., 2003b). 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. (2003b) found that the model-simulated response to combined anthropogenic and natural forcing is robustly detectable in different reanalysis products, even when global mean tropopause height increases are removed. Solar and volcanic forcing alone could not explain the tropopause height increases (Figure 9.4.10). Climate data from re-analyses, especially the “first generation” re-analysis analysed by Santer et al (2003b) are subject to some deficiencies, notably inhomogeneities related to differing 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). To assess the sensitivity of results to such uncertainties and deficiencies, Santer et al (Santer et al., 2004) analysed the “second generation” ERA-40 reanalysis and found that this new dataset also showed a significant anthropogenic influence on tropopause height.

[INSERT FIGURE 9.4.10 HERE]

9.4.4.3 Overall atmospheric temperature change

Satellite-borne microwave sounders (MSU), beginning in 1978, estimate the temperature of thick layers of the atmosphere. The main layers are discussed in Chapter 3, Section 3.4.1.2.1 and represent the lower troposphere (T2LT), the mid troposphere (T2) and the lower stratosphere (T4). Santer et al. (2003c) compared T2 and T4 temeperature 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, which are discussed in 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.

A robust detection of anthropogenic influence on free atmosphere temperatures has been found in a number of different studies analysing various versions of the HadRT.ls radiosonde dataset (Parker et al., 1997) by means of a variety of different diagnostics and fingerprints estimated by the HadCM2 and HadCM3 models (Tett et al., 2002; Thorne et al., 2002; Jones et al., 2003; Thorne et al., 2005a). These studies also found that the magnitudes of tropospheric warming in these models are larger than in the radiosonde based records, although these records could contain spurious cooling trends (Sherwood et al., 2005; Randel and Wu, 2006; Chapter 3, Section 3.4.1). 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 robustly detecting the effects of greenhouse gases and sulphate aerosols (Jones et al., 2003). 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 both models analysed, HadCM2 and HadCM3, over-estimate free-atmosphere warming. Douglass et al. (2004) also concluded that models predicted stronger mid- and upper-tropospheric warming than had been observed. However, like many other studies, Douglass et al did not take account of observational uncertainty. Structual uncertatinty, arising from the choice of techniques used to analyse radiosonde data has not yet been quantified (Thorne et al., 2005b) and some of the disagreement between modelled and observed warming trends could be due to spurious cooling trends in radiosonde records (Sherwood et al., 2005; Chapter 3, Section 3.4.1).

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. They showed that the HadCM3 model appears to underestimate the observed free-atmosphere response to solar forcing, which was also seen by Stott et al. (2003a) for near-surface temperatures, presumably because HadCM3 has a lower sensitivity to solar forcing than greenhouse forcing (Gregory et al., 2004b; see Section 9.4.1.4).

A different approach is to compare observed temperature changes through the depth of the atmosphere with atmosphere-only general circulation model simulations forced with observed sea surface temperatures. The vertical profile of the atmosphere 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) showed that it is highly unlikely that the observed changes could be accounted for by sea surface temperature variations and internal variability alone. They found that inclusion of anthropogenic effects improved the simulation of zonally averaged upper air temperature changes such that an anthropogenic signal was detected at the 5% significance level on both 8-year and inter-annual timescales.

9.4.4.4 Differential temperature trends

Differencing temperature trends at the surface and 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 (CCSP, 2006). Since 1979, globally averaged modelled trends in topospheric lapse rates are consistent with those observed. However, in the tropics most models have more warming aloft than at the surface while most observational estimates show more warming at the surface than in the troposphere (CCSP, 2006). CCSP (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 accounted for (CCSP, 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 likely 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; CCSP, 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; CCSP, 2006) .

A systematic intercomparison between radiosonde based (RATPAC, Free et al., 2005, and HadAT, Thorne et al., 2005b) and satellite based (RSS, UAH) observational estimates of tropical lapse rate trends with those simulated by 19 IPCC AR4 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; CCSP, 2006). Whilst Vinnikov et al. (2006) have not produced a lower tropospheric retrieval, their T2 estimate of temperature trends is consistent with model simulations (Chapter 3, Figure 3.4.3; CCSP, 2006). It is only on longer timescales that disagreement between modelled and observed lapse rates arises, the timescales on which discrepancies arising from inhomogeneities in the observational record would be expected to become apparent. Only one observational dataset (RSS) is consistent with the models on both short and long timescales. One possible explanation 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 remaining errors that affect their long term trends (Chapter 3, Section 3.4.1; CCSP, 2006).

9.4.5 Summary

Since the TAR, the evidence has strengthened that global temperatures have increased near the surface of the Earth as a result of human influence. 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 climate models are only able to reproduce observed temperature changes over the 20th century when they include anthropogenic forcings and their failure to do so when they exclude anthropogenic forcings is strong evidence for the influence of humans on global climate. 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. This conclusion is robust to details of model formulation and to uncertainties in forcings; as far as they have been explored in the large multi-model ensembles available (see Figure 9.4.1).

Many studies have detected a human influence on near-surface temperature changes, applying a variety of statistical techniques and using many different climate model simulations. Comparison with observations shows that these models appear to have an adequate representation of internal variability on the decadal to inter-decadal time-scales important for detection (Figure 9.4.3). When framed in a Bayesian framework, evidence for a human influence on global temperature change is found to be “very strong”, 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. Estimates obtained from optimal detection analyses applied to spatial and temporal changes in temperature over the 20th century indicate that there has likely been a cooling influence from aerosols and natural forcings counter-acting some of the warming influence of the increasing concentrations of greenhouse gase concentrations (Figure 9.4.5). 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 hemisphere 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. Analyses based solely on temporal changes in global mean temperatures do not consistently detect the cooling effects of aerosols and Bayesian studies, which find very strong evidence for an anthropogenic influence on climate, find weaker evidence for a significant effect of aerosols. Nevertheless, the accumulated evidence from a variety of optimal detection studies applied to a wide range of different model simulations (Figure 9.4.5) shows a significant aerosol cooling influence over the last five decades. Therefore, despite continuing uncertainties in aerosol forcing and the climate’s response, it is likely that greenhouse gases alone would have caused more than the observed warming over the last fifty years.

An important development since the TAR has been the identification of an emerging anthropogenic signal in surface temperature changes on continental and sub-continental scale land areas. The ability of models to simulate many aspects of the temperature evolution on these scales (Figure 9.4.8) and the detection of significant anthropogenic effects on individual continents (Figure 9.4.7) provide compelling evidence for human influence on climate. Although it is generally more difficult to attribute temperature changes to individual forcings in continental and sub-continental regions, and in individual seasons than to attribute global scale changes, human influence has been detected in a variety of different studies in a number of disparate regions. The chance that all regional results in different parts of the globe are spurious is very small, particularly considering that different regions are affected by different uncertainties in observations, external forcings and internal variability.

Evidence for changes in extreme temperatures is beginning to emerge. There has been a significant decrease in the frequency of frost days and an increase in the incidence of warm nights. An optimal detection 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 9.4.9). Changes in the probability and recurrence time of extreme temperatures, as well as extreme rainfall and storminess, are expected under climate change conditions, suggesting that this kind of quantitative risk analysis will become more important in the future.

Since the TAR further evidence has accumulated that there has been a significant anthropogenic influence on free atmosphere temperature since measurement 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 9.4.10).

Simulations of differential warming rates between the surface and the free atmosphere are inconsistent with some observational records. These possible discrepancies between modelled and observed lapse rates are largest in the tropics, where interpretation is critically dependent upon the observationally derived dataset used in the comparison. However, the IPCC AR4 simulations are remarkably consistent in their predictions of tropical tropospheric changes despite differences in models and forcings. Both models and observations do show slow decadal changes in lapse rate (both positive and negative) particularly over the longer period of the radiosonde era. Further understanding of lapse rate changes will require a more systematic treatment of the full range of both modelling and observational uncertainty.

9.5 Understanding of Change in Other Variables during the Industrial Era

The objective in 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 state changes

Warming of the surface should lead, in time, to warming of the sub-surface ocean, and hence, due to thermal expansion, an increase in sea level. Melting of glaciers and ice caps will also contribute to an increase in sea level. Since the TAR there has been an accumulation of evidence for climate change within the ocean, both at regional and global scales (see Chapter 5). The overall heat content in the world ocean is estimated to have increased by 14.1×10^{22} J during the period 1961–2003 (see Chapter 5, Section 5.2.2, Figure 5.2.5, and Table 5.2.1). This overall increase has been superimposed on strong interdecadal variations. The fact that the ocean gained heat during the last 50 years is in itself a strong argument for a net positive radiative forcing of the climate system. If the observed warming of the atmosphere (see Chapter 3; Section 9.4) originated from natural internal sources of variability, then the source of the heat would have likely been the ocean given that it is by far the system's largest heat reservoir (Levitus et al., 2005; Chapter 5, Figure 5.2.5). However, observations indicate that the ocean has been gaining, rather than losing, heat, suggesting an external rather than internal heat source. For a consistent explanation of observed changes, models should be able to simulate both the overall increase and capture the observed variability.

9.5.1.2 Heat content in ocean basins

Levitus et al. (2000) presented evidence that there had been a warming of the world ocean in the second half of the 20th century (with evidence from geologic sources or earlier increases in ocean heat content; Section 9.3; Crowley et al., 2003). These estimates were subsequently updated to include additional data both for earlier and more recent years (Levitus et al., 2005) which resulted in a small revision downwards of the observed increase in ocean heat content (see Chapter 5, Section 5.2.2 and Figure 5.2.5).

A number of studies have sought to understand the late 20th century changes, which Levitus et al. (2001) demonstrated were at least one order of magnitude larger than the increase in heat content of any other component of the earth's ocean-atmosphere-cryosphere system. Levitus et al. (2001) and Gregory et al (2004a) analysed simulations of the GFDL R30 and HadCM3 models respectively and showed that model 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. (2002) used optimal 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. All these analyses indicate a large anthropogenic component to the positive trend in global ocean heat content. 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)**.** It is likely that the cooling influence of natural (volcanic) and anthropogenic aerosols has slowed ocean warming; Delworth et al. (2005) find a delay of several decades and a reduction in magnitude of warming of approximately two thirds in simulations of the GFDL CM2 model including these forcings when compared to the response to increasing greenhouse gases alone, consistent with results based on an upwelling diffusion energy balance model (Crowley et al., 2003). Important reductions in ocean heat content are found following volcanic eruptions in model 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). Barnett et al. (2005) considered the detection implications of the revisions to the Levitus et al. (2000) ocean heat content data (Levitus et al., 2005) and found that the earlier conclusions of Barnett et al. (2001) were not affected.

Barnett et al (2005) extended previous analyses of ocean heat content changes to a basin by basis analysis of the temporal evolution of temperature changes in the upper 700m of the ocean (see also Pierce et al., 2006). They report that whereas natural internal variability and naturally externally forced variability as simulated by the PCM are not capable of replicating the observed signal, simulated ocean warming due to anthropogenic factors (including well mixed greenhouse gases and sulphate aerosols) is consistent with the observed changes and reproduce many of the different responses seen in the individual ocean basins (Figure 9.5.1). Repeating the analysis using the HadCM3 model supported their conclusions of a human induced warming of the world's oceans with a complex vertical and geographical structure that is simulated quite well by two AOGCMs.

[INSERT FIGURE 9.5.1 HERE]

Although the overall increase in ocean heat content is well explained by models, the decadal variability seen in Levitus et al. (2000; 2005) (see Chapter 5, Section 5.2.2) is not well reproduced by models. Gregory et al. (2004a) 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 in-filling the observational dataset outside this well-observed region. They find a strong maximum in variability in the Levitus dataset at around 500m depth that is not seen in HadCM3, a possible indication of model deficiency or alternatively an artefact in the Levitus data. AchutaRao et al. (2005) 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.

9.5.1.3 Water mass properties

The ocean heat content analyses cited above use basin-integrated values for the different ocean basins, in part because observational coverage, particularly at lower levels in the ocean, remains thin. However some studies have attempted to investigate changes in three dimensional water mass properties (see Chapter 5, Section 5.3). Sub Antarctic mode water 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 (see Chapter 5, Section 5.6) These changes are consistent with a global increase in the hydrological cycle over the oceans over the last 50 years with increased precipitation at high latitudes (Wong et al., 1999) and a reduction in the difference between precipitation and evapouration at mid-latitudes (see Chapter 5, Section 5.6). This would suggest that the ocean might integrate rainfall changes to produce detectable salinity changes. Boyer et al. (2005) provided linear trend estimates of salinity for the World ocean from 1955 to 1998, indicating salinification in the Antarctic Polar Frontal Zone around 40S and in the subtropical North Atlantic, and freshening in the sub-polar Atlantic (see Chapter 5, Figures 5.3.1, and 5.2.6).

Care should be taken in interpreting sparse hydrographic data, since apparent trends could be aliased natural variability or the aliased effect of changing observational coverage. One such example concerns recent measurements of the southern Indian Ocean gyre along the WOCE 15 section. Sub Antarctic Mode Water (SAMW) in the South Indian Ocean was fresher on isopycnals in 1987 than in the 1960s, but in 2002 the salinity was again near to the 1960s values (Bindoff and McDougall, 2000; Bryden et al., 2003). Based on 20th Century simulations with the HadCM3 model, it is not possible to reject the null hypothesis that the observed differences are due to internal variability (Stark et al., 2006). However the model suggests a long-term freshening trend in the 21st Century due to the large scale response to surface heating and hydrological changes (Banks et al., 2002).

One possible oceanic consequence of climate change is a slowing down or even halting of the meriodional overturning circulation in the Atlantic (MOC, Chapter 5, Box 5.1). An estimate of the overturning circulation and associated heat transport based on a transatlantic section along latitude 25N indicates that the Atlantic MOC has slowed by about 30% between 1957 and 2004 (Bryden et al., 2005). Freshening of North East Atlantic Deep Water has been observed (Dickson et al., 2002; Chapter 5, Figure 5.3.4; 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 an increasing rather than a decreasing 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; Chapter 5, Box 5.1) and the Atlantic Multidecadal Oscillation (Chapter 5, Box 5.1, Vellinga and Wu, 2004; Knight et al., 2005). In summary, it is not yet possible to attribute changes in MOC properties to natural and anthropogenic causes (Chapter 5, Box 5.1).

9.5.2 Sea level

In order to have confidence in model-based projections of future sea level change resulting from climate change, model-based estimates of historical global mean sea level rise must be consistent with observational estimates. Models also offer the possibility of attributing sea-level changes to particular forcing factors. Results from 15 IPCC AR4 C20C3M simulations are compared to observations (details, see Chapters 4 and 5). The simulations show trends of 0.65 ± 0.50 K century^{-1} in their timeseries of global average surface air temperature change over the 20th century The uncertainty bounds, which represent model uncertainty only, are ±2 standard deviations of the inter-model varation in trend.These surface temperature trends are consistent with observations (Section 9.4.1, Figure 9.4.1), but this does not guarantee a realistic simulation of thermal expansion, as there may be compensating errors among climate sensitivity, ocean heat uptake and radiative forcing (c.f. Raper et al., 2002).The observational budget for sea level (Section 5.5.6) assesses the periods 1961-2003 and 1993-2003, but most C20C3M simulations end earlier (in 1999-2002), so the comparison that follows is not quite exact.

The model range of 0.68 ± 0.54 mm yr^{-1} for simulated thermal expansion for 1961-2003 is somewhat above the observational range of 0.42 ± 0.14 mm yr^{-1} for an estimate of thermal expanstion based on observed ocean warming (Chapter 5, Section 5.5.3). However, model spread is likely enhanced because many models considered do not include volcanic forcing; several large volcanoes cooled the climate during the 1960s, and the models with this effect generally have smaller ocean heat uptake in recent decades (Gleckler et al., 2006). mean of models using anthropogenic and natural forcing is 0.57 mm yr^{-1} (model IDs 3, 11, 12, 14, 15 and 21; see Chapter 8, Table 8.2.1), and of those anthropogenic forcing only is 0.76 mm yr^{-1} (model IDs 4, 6, 7, 8, 13, 16, 18, 19, 20 and 22; see Chapter 8, Table 8.2.1). For 1993-2003 the observational range is 1.6 ± 0.6 mm yr^{-1} and the model range is 1.2 ± 1.0 mm yr^{-1}which agrees with the thermal expansion of 1.2 ± 0.6 mm yr^{-1} calculated by Antonov et al. (2005) and Ishii et al. (2006) for the upper 700 m of the ocean (see discussion in Chapter 5). Note that the estimated thermal expansion for this period from models using natural and anthropogenic forcing is 1.5 mm yr^{-1} while that estimated from models using only anthropogenic forcing is 1.0 mm yr^{-1}. Although this is uncertain because these simulations end at various dates from 1999 onwards, it agrees well with a result obtained by Church et al. (2005) using the PCM, which suggests 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. This is supported by Gregory et al. (2006) using HadCM3. The large model spread is probably caused by fact that only some models include Pinatubo, and possibly also that many runs end before 2003.

Volcanoes can influence the ocean on the shorter and longer periods. For example, after 1993 the short term response is determined by the recovery of the upper ocean following cooling due to the eruption of Pinatubo in 1991 whereas the multi-decadal response is related to the much longer persistence of cool anomalies in the deep ocean (Gleckler et al., 2006). For both 1961–2003 and 1993–2003, a discrepancy between model and observations could also be partly explained by the internally generated variability of the climate system, which model control runs suggest could give a standard deviation of ~0.2 mm yr^{-1} in 10-year trends and ~0.1 mm yr^{-1} in 40-year trends; the latter in particular may be underestimated by models (see Chapter 5, Section 5.2.2.2; Section 9.5.1.2; Gregory et al., 2006).

Using the AOGCM results and the glacier mass balance model of Oerlemans (2001) and Oerlemans et al., (2005) (see also Chapter 10, Section 10.6.3.1) suggests a sensitivity of global glacier and ice caps balance (excluding Greenland and Antarctica) to global temperature change of 0.5 ± 0.2 mm yr^{-1} K^{-1}, consistent with 0.65 ± 0.40 mm yr^{-1} K^{-1} from a regression of glacier mass balance observations (see Chapter 4, Section 4.5.2) against global temperature up the mid-1990s (the correlation is 0.8). Adopting a value of 0.65 mm yr^{-1} K^{-1} and assuming the climate of 1900–1929 to be 0.15 K warmer than the steady state for glaciers (see discussion in Chapter 10, Section 10.6.3), the AOGCMs give 0.4 ± 0.2 mm yr^{-1} of sea-level rise from glacier and ice caps in 1961–2003 and 0.6 ± 0.3 mm yr^{-1} in 1993 ± 2003, similar to the observational estimates of 0.49 ± 0.33 mm yr^{-1} and 0.78 ± 0.46 mm yr^{-1} respectively respectively (see Chapter 5, Section 5.5.5.2). Accelerated global glacier and ice cap mass loss in more recent years may indicate a greater sensitivity.

Calculations of ice-sheet surface mass balance changes due to climate change (following the methods of Gregory and Huybrechts, 2006, and Chapter 10, Section 10.6.3.1) indicate probably small but uncertain contributions during 1993–2003 of 0.1 ± 0.2 mm yr^{-1} 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). The observational estimates (see Chapter 4, Section 4.6.2 and Chapter 5, Section 5.5.6) are 0.21 ± 0.07 mm yr^{-1} and 0.21 (sic) ± 0.35 mm yr^{-1} respectively. 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, Section 10.6.4 for discussion). In addition Antarctica may be contributing 0.0–0.2 mm yr^{-1} on account of its ongoing adjustment to the end of the last glacial period (see Chapter 6, Section 6.5). These dynamic terms are sufficient to explain the difference between model and observations.

Adding the thermal expansion and global glacier and ice cap terms (excluding the ice sheets) we obtain 1.1 ± 0.7 mm yr^{-1} for 1961–2003 and 1.8 ± 1.0 mm yr^{-1} for 1993–2003, lying below the observed rates of 1.8 ± 0.5 mm yr^{-1} and 3.1 ± 0.8 mm yr^{-1} (see Chapter 5, Sections 5.5.1.1 and 5.5.1.2). For the earlier period, the terms are reasonably well reproduced by the models, and the discrepancy between many models and observations indicates the lack of a satisfactory explanation of sea-level rise (discussed in Chapter 5, Section 5.5.6). In the later period, the ice sheets contribute 0.4 ± 0.4 mm yr^{-1}. The remaining discrepancy The remaining discrepancy could partly be because of deficient model variability, since tide gauge estimates suggest that rates of sea level rise as large as during 1993–2003 have occurred in earlier decades (see Chapter 5, Section 5.5.2.1.3; see also Gregory et al., 2006).

An increase in the rate of rise over recent decades could have been caused by rising anthropogenic forcing (Woodworth et al., 2004), which is expected to have been the largest contributor to the overall sea level rise during the 20th century (Crowley et al., 2003; Gregory et al., 2006). The models we have analysed here show accelerations during the 20th century of 0.014 to 0.009 mm yr^{-2}, consistent with the observational estimate of 0.012 ± 0.006 mm yr^{2} made by Church and White (2006). On the other hand, natural forcings in the early 20th century could have partially offset the increasing anthropogenic forcing and tended to produce a steadier rate of rise during the 20th century (Crowley et al., 2003; Gregory et al., 2006). The rate of sea level rise was greater in the 20th than in the 19th century (see Chapter 5, Section 5.5.2.5). 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 become important later in the 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) (see Chapter 3, Section 3.6). The extent to which these modes can be excited or altered by external forcing remains uncertain, but their impact on terrestrial climate on annual to decadal time scales can be profound. While at least some of these modes can 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 based on synoptic diagnostics (Risbey and Kandlikar, 2002).

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 (see Chapter 3, Section 3.6.2). There have been multidecadal oscillations in the ENSO index throughout the 20th century, with more intense El Niño events since the late 1970s, partly corresponding to a mean warming of the western equatorial Pacific (Mendelssohn et al., 2005). The 1998 El Niño was the strongest on record. 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, 2000a), others have shown no change (e.g., Collins, 2000b), or a decrease in variability (Knutson et al., 1997). A recent survey of the simulated response to CO_2 doubling in fifteen IPCC AR4 coupled climate models (Merryfield, 2006) found that three of the models exhibited significant increases in ENSO variability, five exhibited significant decreases and seven exhibited no significant change. Using a model which simulated an increase in variability, Timmerman (1999) found no detectable change in ENSO variability in the observations. 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.

Decadal variability in the North Pacific is characterised by variations in the strength of the Aleutian Low coupled to changes in North Pacific SST (see Chapter 3, Section 3.6.3). 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 (Latif and Barnett, 1994; Mantua et al., 1997; Zhang et al., 1997; Deser et al., 2004). However, the time-series corresponding to the PDO exhibits more variability at decadal and longer time-scales than traditional ENSO SLP or SST derived indices (Mantua et al., 1997; Newman et al., 2003). Recent work suggests that the PDO may be the North Pacific expression of an ENSO-like pattern of variability called the Interdecadal Pacific Oscillation or IPO (Folland et al., 2002; Deser et al., 2004). It is presently not clear to what extent the PDO or IPO is physically different from tropical Pacific variability in El Niño, and also to what extent extratropical influences play a role (e.g., Liu et al., 2002; Newman et al., 2003; Wu et al., 2003).

The PDO/IPO is an internally generated mode of climate variability associated with variability that shows large-scale influences in both models and observations (e.g., Pierce et al., 2000; Yukimoto et al., 2000; Salinger et al., 2001; Arblaster et al., 2002; Reason and Roualt, 2002; Vimont et al., 2002; Deser et al., 2004; Shiogama et al., 2005). Like ENSO, the PDO/IPO influences hemispheric and global average surface temperature (Pan and Oort, 1983; Meehl et al., 1998; Bratcher and Giese, 2002), as well as many aspects of regional climate, such as Amazonian rainfall (Marengo, 2004) and Northern Hemisphere terrestrial rainfall (Deser et al., 2004). One recent study identified an anthropogenic influence towards the positive phase of the PDO based on simulations of the MIROC model (Shiogama et al., 2005).

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 (see 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 for a detailed discussion; 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., 2002c; 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., 2002c; Gillett et al., 2003b; Osborn, 2004; Gillett, 2005; and see Figure 9.5.2). 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., 2002c; 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 Northern Annular Mode. Stenchikov et al. (2006) examined changes in sea level pressure following nine volcanic eruptions in 20th century (20C3M) simulations of the IPCC AR4 ensemble, and found that the models simulated a positive NAM response to the volcanoes, albeit one that was smaller than that observed. However, 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., 2003b). Simulations incorporating all the major anthropogenic and natural forcings from the IPCC AR4 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 9.5.2), 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).

[INSERT FIGURE 9.5.2 HERE]

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., 2001; Hurrell et al., 2005), suggesting that the trend may in part relate to SST changes, particularly over the Indian Ocean (Hoerling et al., 2005b). 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 tropical 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, Figure 3.6.6). 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 9.5.2). 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. By contrast, Jones and Widmann (2004) develop a 95-year reconstruction of the summer SAM index based largely on mid-latitude pressure measurements, and find that the recent upward trend in the SAM is not unprecedented, suggesting it may have a natural cause. However, reconstructions from 1958 using additional data indicate 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, 2005; 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) , probably by changes to the propagation of planetary waves, although radiative effects may also play a role (Solomon et al., 2005). 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; Arblaster and Meehl, 2005; Stone and Ffye, 2005) . 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 Antarctic plateau cooling , and about a third to half part of the warming of the Antarctic Peninsula (Thompson and Solomon, 2002; Carril et al., 2005; Chapter 3, Section 3.6.5), 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

Trends in sea level pressure may be assessed globally by applying optimal fingerprinting techniques similar to those which have been applied to temperature. 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 a detection and attribution analysis using 20C3M simulations of eight IPCC AR4 coupled models (Gillett et al., 2005) (Figure 9.5.2). However, 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. By contrast in the Northern Hemisphere simulated sea level pressure trends are much smaller than those observed (Gillett, 2005).

[INSERT FIGURE 9.5.2 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 (see Chapter 3, Section 3.7.1 and Section 9.5.3.2). The Asian monsoon circulation in the IPCC AR4 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 (see Chapter 3, Section 3.7.1). These results are consistent with the expectation (Ramanathan, 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 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 3.4 and Trenberth, 2005). The spatial resolution of most climate models limits their ability to realistically simulate tropical cyclones (see 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 SSTs, or embedded hurricane models in lower resolution GCMs (see Chapter 10, Section 10.3.6.3). While results vary, 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 (see Chapeter 10, Section 10.3.6.3). It has been suggested that the simulated reduction in frequency of tropical cyclones under enhanced CO_2 conditions may result from an increase in radiative cooling associated with increased CO_2 concentration (Sugi and Yoshimura, 2004; Yoshimura and Sugi, 2005; Chapter 10, Section 10.3.6.3), 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; Chapter 3, Section 3.8.3.1). Positive trends in tropical cyclone potential intensity (a measure of the maximum intensity expected of tropical systems) are found in some tropical locations (Bister and Emanuel, 2002) but other studies find no consistent trends (Free et al., 2004; Chapter 3, Section 3.8.3.1). Emanuel (2005) reports a marked increase since the mid-1970s of an index of the intensity of tropical cyclones (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. This index is closely related to tropical sea surface temperatures, including oscillations such as ENSO, the NAO, and the Atlantic Multi-decadal Oscillation (AMO – Chapter 3, Section 3.6.6.1), as well as the strong tropical mean warming since the mid-1970s, suggesting that the increased intensity may be partly the result of warming. The changes appear to have been the result of increases in both the duration of cyclones and their peak intensity. Webster et al. (2005) found a strong increase in the number and proportion of the most intense tropical cyclones over the past 35 years. Nonetheless, detection and attribution of observed changes in hurricane intensity or frequency to external influences remains difficult given deficiencies in theoretical understanding of tropical cyclones, their modelling, and their long-term monitoring.

9.5.3.7 Extra-tropical cyclones

Simulations in the IPCC AR4 20C3M model ensemble generally show a decrease in the total number of extratropical cyclones in both hemispheres, but an increase in the number of the most intense events when compared to pre-industrial control simulations (Lambert and Fyfe, 2006). Similar changes were seen in earlier model studies (Fyfe, 2003; Geng and Sugi, 2003), and in the ERA-40 reanalysis, though these are not statistically significant.

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 (see 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 (see Chapter 3, Section 3.5.3).

9.5.4 Precipitation

9.5.4.1 Changes in atmospheric water vapour

As climate warms, the amount of moisture in the atmosphere is expected to rise (Trenberth et al., 2005), since 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 0.40 ± 0.09 mm per decade averaged over the global oceans (see Chapter 3, Section 3.4.2.2). Soden et al. (2005) demonstrated that these observed changes are well simulated in the GFDL atmosphere model with prescribed SSTs (Figure 9.5.3), including the upward trend. 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 their model.

[INSERT FIGURE 9.5.3 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. However, global mean precipitation is controlled primarily by the energy budget of the troposphere, which is in turn influenced by greenhouse gas and aerosol forcing, and the net anthropogenic response remains unclear (Allen and Ingram, 2002). Global annual land mean precipitation showed a small, but uncertain, upward trend over the 20th century of approximately 1.1 mm/decade (see Chapter 3, Section 3.3.2.1 and Table 3.4). However, the record is characterised by large interdecadal variability, and global terrestrial annual mean precipitation shows almost no trend since 1940 (Figure 9.5.4; see also Chapter 3, Table 3.4).

9.5.4.2.1 Detection of external influence on precipitation

While greenhouse gas increases are expected to cause an increase in precipitation over most of the globe, enhanced horizontal transport and changes in moisture convergence is expected to lead to a pronounced drying of the subtropics, and subsiding regions of the tropics, while contributing a further increase in precipitation in the equatorial region and at high latitudes (Emori and Brown, 2005; Neelin et al., 2006). However, on regional scales model agreement in predicted change is quite poor, and the change expected in the tropics thus far is quite small (Neelin et al., 2006).

Mitchell et al. (1987) argue that 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 the cooling due to a CO_2 increase (Allen and Ingram, 2002; Yang et al., 2003; Lambert et al., 2004; Sugi and Yoshimura, 2004). Yang et al. (2003) and Sugi and Yoshimura (2004) demonstrated that in a climate model with doubled CO_2 but fixed sea surface temperatures, the rate of radiative cooling is decreased by the increased CO_2, while the temperature of the atmosphere remains constrained by the fixed lower boundary, leading to a reduction in precipitation. This mechanism therefore suggests that precipitation should respond more to changes in shortwave forcing than longwave 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 simulations of the response to volcanic forcing which indicate that global precipitation decreases by approximately 1 to 3% following large volcanic eruptions (Robock and Liu, 1994; Broccoli et al., 2003). It also explains why anthropogenic influence is not yet detectable in measurements of global land mean precipitation (Ziegler et al., 2003; Gillett et al., 2004b). However, Lambert et al. (2004) urge caution in applying the energy budget argument to land-only data, where they argue that the availability of moisture is in many cases more important than the energy budget of the troposphere for controlling precipitation. 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 terrestrial precipitation in climate model integrations (see Chapter 8, Section 8.3.1.2 for an evaluation of model simulated precipitation) 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. Lambert et al. (2005) examine precipitation changes in simulations of nine IPCC AR4 20C3M models including anthropogenic and natural forcing (Figure 9.5.4a). Using five of the models they find a detectable response to external forcing in the observations, although in two cases the residual variance is unrealistically large. Lambert et al. (2004) further demonstrate using HadCM3 that the response to shortwave, but not longwave forcing is detectable in land mean observations. Gillett et al. (2004b) similarly demonstrate that the terrestrial precipitation response to volcanic forcing simulated by the PCM is detectable in observations. These results are therefore consistent with expectations that natural forcings such as volcanic aerosol are likely to have had a larger influence on land mean precipitation than have greenhouse gas changes thus far. Lambert et al. (2005) report that all nine models they examine underestimate the variance of terrestrial precipitation compared to that observed, consistent with earlier findings (Gillett et al., 2004b; Lambert et al., 2004). It is unclear whether this discrepancy results principally from an underestimated response to shortwave forcing (Gillett et al., 2004b), underestimated internal variability, or errors in the observations.

[INSERT FIGURE 9.5.4 HERE]

While greenhouse gas increases are expected to cause an increase in precipitation over most of the globe, enhanced horizontal transport is expected to lead to a pronounced drying of the subtropics, while contributing a further increase in precipitation in the equatorial region and at high latitudes (Emori and Brown, 2005). Thus simulations of twentieth century zonal mean land precipitation generally show an increase at high latitudes and at the equator, with a pronounced decrease in the subtropics of the Northern Hemisphere (Figure 9.5.4b; Hulme et al., 1998), and simulations of the 21st century show a similar effect (see Chapter 10, Figure 10.3.9). 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 9.5.4b; Hulme et al., 1998; Rotstayn and Lohmann, 2002), although the observed record is characterized by large interdecadal variability (see Chapter 3, Figure 3.3.4). This agreement between the simulated and observed zonal mean precipitation trends is not found to be sensitive to the inclusion of volcanos in the simulations, suggesting that a response to anthropogenic influence may be emerging in this diagnostic.

Changes in streamflow have been observed in many parts of the world, with increases or decreases corresponding to changes in precipitation (see Chapter 3, Section 3.3.4). Climate models suggest that streamflow will increase in regions where precipitation increases faster than evaporation, such as at high Northern latitudes (see Chapter 10, Section 10.3.2.3 and Figure 10.3.9; see also Milly et al., 2005). Gedney et al. (2005) attributed increased continental runoff in the latter decades of the 20th century 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 run-off. Nonetheless, 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 AR4 models are significantly correlated with observed runoff trends.

Mid-latitude summer drying is another anticipated response to greenhouse gas forcing (see 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 (see 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 that observed and the relative contributions of natural external forcings and anthropogenic forcings was not assessed. The model also simulated some aspects of the regional pattern of drought trends, such as the observed strong 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 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. This Clausius-Clapeyron constraint was found to apply well in HadCM3. 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 in explaining the general increase in extreme precipitation that is seen in models. Meehl et al. (2005), while not specifically concerned with extreme precipitation, demonstrate that tropical precipitation intensity increases due to increases in water vapour, 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 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 (IPCC, 2001; Semenov and Bengtsson, 2002; Trenberth et al., 2003). Trenberth et al. (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 (see Chapter 8, Section 8.5.2 for an evaluation of model simulated precipitation extremes) also suggest that future changes in precipitation extremes will likely be greater than changes in mean precipitation (see Chapter 10, Section 10.3.6.1). Simulated changes in globally averaged annual mean and extreme precipitation appear to be quite consistent between models. A model-model detection study, in which fingerprints from one model were used to detect precipitation change in simulations from another model, suggests that changes in heavy precipitation (i.e., the magnitude of events that occur a few times per year) may be more robustly detectable using signals from different models than changes in annual total rainfall (Hegerl et al., 2004). This is mainly because precipitation extremes increase over a large fraction of the globe in both models, whereas total precipitation exhibits a more model-dependent spatial pattern of increases and decreases.

Evidence for changes in observations of short-duration precipitation extremes vary with the region considered (Alexander et al., 2006) and the analysis method that is employed (IPCC, 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 was made based on Frich et al. (2002) indices, using fingerprints from atmospheric model simulations with prescribed sea surface temperature and a bootstrap method for significance testing (Kiktev et al., 2003). This study indicated that patterns of simulated and observed rainfall extremes bear little similarity. 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 GCMs run for the IPCC AR4 showed a general tendancy towards greater frequency of heavy-precipitation events over the past four decades, most coherently in the high latitudes of the Northern hemisphere.

9.5.4.3 Regional precipitation changes

Regional trends in precipitation are likely to exhibit strong spatial variations because of the dependence of precipitation on atmospheric circulation and on geographic factors such as orography. Trends in observed annual precipitation during the period 1901 to 2003 are shown in Chapter 3, Figure 3.3.2 for regions in which data is available. Anthropogenic influence has not been demonstrated in global patterns of precipitation change, although there is some agreement between observed and simulated changes in zonal mean land precipitation (see Section 9.5.3.2.1 and Figure 9.5.4b).. Also, there have been some suggestions, for specific regions, of an anthropogenic influence on precipitation, which we discuss 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 9.5.5, see also Chapter 3, Figure 3.7.4). 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 three 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, large-scale atmospheric circulation changes related to global sea surface temperature changes, and internal variability (Nicholson, 2001). Black carbon has also been suggested as a contributor (Menon et al., 2002b). 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 Sahel 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; Hoerling et al., 2005a; see also Figure 9.5.5), consistent with earlier findings (Folland, 1986; Rowell, 1996). These studies differ somewhat in terms of which ocean SSTs they find to be most important: Gianniniet al. (2003) and Bader and Latif (2003) emphasize the role of tropical Indian Ocean warming, Hoerling et al. (2005a) 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. (2005a) concluded that the observed drying trend in the Sahel is not consistent with simulated internal variability alone or with the simulated response to greenhouse gas forcing alone.

[INSERT FIGURE 9.5.5 HERE]

Thus recent research indicates that differences in warming of sea surface temperatures in different basins are likely to have been the dominant influence on rainfall in the Sahel, although land use changes possibly contribute also (Taylor et al., 2002). But what has caused the differential changes in sea surface temperatures? 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 model experiment 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 therefore 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 attribution analysis has yet to be applied. Held et al. (2005) showed that historical coupled 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 rainfall also shows some agreement with observations. However, the Held et al. results are in contrast to other coupled model simulations, which generally show little change or even an increase in Sahel rainfall in response to 20th century anthropogenic forcing (Hoerling et al., 2005a), suggesting that the observed late 20th century decline in rainfall is unlikely to be due to this cause.

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). Rainfall has 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 increases in greenhouse gas concentration. 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 the enhanced greenhouse effect likely contributed to the rainfall decrease. Timbal et al. (2005) detected an anthropogenic influence in South Western (SW) Australian precipitation, using climate change signals that were downscaled from the PCM..

Some authors have suggested that the decrease in rainfall is related to changes in the Southern Annular Mode (SAM), with a decrease in SW Australian rainfall associated with a southward shift in the storm track, and anomalously high mid-latitude atmospheric pressure (e.g., Karoly, 2003). Several modelling studies have shown that this change in the SAM may be a response to stratospheric ozone depletion, greenhouse gas increases, or a combination of both (see Section 9.5.3.3). However, 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). Thus it remains unclear how important this circulation trend has been for the Southwest Australian drought.

9.5.4.3.3 Monsoon precipitation

Observations show that overall South Asian, North African and North American monsoon precipitation decreased somewhat in the mid-1970s (see Chapter 3, Section 3.7.1). The TAR (IPCC, 2001, pp 568) concluded that an increase in SE 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). However, by including the effects of black carbon aerosol, Ramanathan et al. (2005) were able to simulate realistic changes in Indian monsoon rainfall, particularly a decrease which occurred between 1950 and 1970. 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. These changes in the monsoon were not reproduced in simulations with greenhouse gas and sulphate aerosol changes only.

9.5.5 Cryosphere Changes

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.

9.5.5.1 Sea ice

The annual mean area of Arctic sea ice has decreased in recent decades, with stronger declines in summertime than winter and some thinning of the ice (see Chapter 4, Section 4.4). Gregory et al. (2002b) in a four-member ensemble of integrations of HadCM3 with forcings due to all major anthropogenic and natural climate 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. Gregory et al. (2002b) concluded that internal variability and natural forcings (solar and volcanic) are “very unlikely by themselves to have caused a trend of this size”. Johannessen et al. (2004) conducted similar integrations with the ECHAM4 model and concluded “there are strong indications that neither the Arctic warming trend nor the decrease in ice extent and volume over the last two decades can be explained by natural processes alone”. 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. Model simulations of historical Arctic ice thickness or volume (Rothrock et al., 2003 and Goeberle and Gerdes, 2003) simulate 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. However, most models indicate a maximum in ice thickness in the mid 1960s, with local maxima around 1980 and 1990 as well. There is an emerging indication from both models and observations that much of the change in thickness occurred between the late 1980s and late 1990s. Although some of the dramatic change inferred may be a consequence of 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 it’s thinning, appears to be largely, but not wholly, due to warming.

Unlike the Arctic, the Antarctic has not exhibited a strong decline in sea ice extent, at least during the period of satellite observations. 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 (over 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 20th century simulations from IPCC models. Many of the models simulate dipole-like behaviour in sea ice extent, which, consistent with observations, are led by sea-level pressure variations in the Amundsen/Bellingshausen Sea. The models exhibit a wide variety of long-term variations through the 20th century, including the period (post 1978) with satellite observations of sea-ice extent. Only one model simulated a positive trend in the period 1979–2000, as was observed. Shindell and Schmidt (2004), using the GISS model forced by ozone and greenhouse gas changes, simulated a December-May cooling of Antarctica over the period 1979–2000, more consistent with the observations of temperature and the lack of a decline in sea ice extent.

9.5.5.2 Snow

Snow cover in the Northern Hemisphere, as measured from satellites, has declined substantially in the past 30 years, particularly from early spring through summer (see Chapter 4, Section 4.2). Trends in snow depth and cover can be driven by precipitation or temperature trends, and generally the trends in recent decades have 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 (equivalent or 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 statistically demonstrated 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.

9.5.5.3 Alpine and polar glaciers

During the 20th century, glaciers have generally experienced considerable mass losses with strongest retreats in the 1930s and 1940s and after 1990 (see Chapter 4, Section 4.5). The widespread shrinkage appears to imply widespread warming as the likely cause (Oerlemanns, 2005), and since glacier retreat on the century time scale is rather uniform across the globe, this suggests that the warming is a wide spread cause although in the tropics changes in atmospheric moisture might be contributing (see 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. (2002) forced a glacier mass balance model for the Nigardsbreen and Rhone 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 likely to be due to externally forced climate change. As well, the thinning and accelerating of at least some polar glaciers appears to be the result of ocean warming (Thomas et al., 2004). The increasing rate of glacier retreat in the Antarctic Peninsula is broadly compatible with retreat driven by warming, but the speed of the retreat suggests that other factors may be contributing.

9.5.5.4 Polar ice sheets and ice shelves

The ice sheets appear to be near-balance or thickening slightly in central regions, but thinning around the margins of Greenland and important parts of West Antarctica, broadly consistent with expectations for a warming world (see Chapter 4, Section 4.6). Warming increases low-altitude melting and high-altitude precipitation in Greenland, with sufficient warming leading to dominance of the melting. 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 (see Chapter 4, Section 4.6.3.2). As well, recent ice-shelf changes have likely resulted from warming of both oceanic and atmospheric temperatures (see Chapter 4, Section 4.6.3.4).

9.5.5.5 Frozen ground

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 (see Chapter 4, Section 4.7.2). Thickness of seasonally-frozen ground has decreased in response to winter warming and increases in snow depth (see Chapter 4, Section 4.7.3).

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 further strengthening the case for an anthropogenic influence on climate, and improving our 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.

ENSO has not shown any behaviour that is clearly distinguishable from natural variability, but both the Northern and Southern Annular Modes have shown significant trends. While models reproduce the sign of the Northern Annular Mode trend, the simulated response is too small. By contrast, models including both greenhouse gas and ozone simulate a realistic trend in the Southern Annular Mode, leading to a detectable human influence on global sea level pressure. Anthropogenic influence is not, as yet, detectable 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 led 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 yet 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 intensity 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, although Southern Hemisphere sea ice extent has not declined as predicted in one model. There is very likely a decreasing trend in global snow cover, and widespread melting of glaciers, consistent with a widespread warming.

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 response* (ECS) is defined as the equilibrium annual global mean temperature response to a doubling of equivalent CO_2 from preindustrial levels and *transient climate response* (TCR) is the annual global mean 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). While the radiative forcing that corresponds to greenhouse gas increases and the direct temperature change that results can be calculated in a relatively straightforward manner (see Chapter 2), feedbacks to radiative forcings are very uncertain (see Chapter 8), leading to uncertainties in estimates of future climate change. The objective here is to assess estimates of climate sensitivity that use observed climate changes to attempt to quantifying and reducing those uncertainties, while Chapter 8 reports direct assessments of feebacks. 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 typically associated with substantially different climates than that of the present, such as the climate of the Last Glacial Maximum. 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. Thus a directly derived estimate of the sensitivity 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 their equilibrium climate sensitivity, and attaching probabilities to different values for the equilibrium climate sensitivity based on the realism of these simulations. This circumvents the problem of feedbacks being dependent on the climatic state at least to the extent that models realistically account for these changes.

Estimates of equilibrium climate sensitivity or transient climate response discussed here are generally based on large ensembles of climate model simulations in which uncertain parameters influencing the model's sensitivity to forcing are varied. Differences between simulated temperature *changes* and those observed in either instrumental or proxy data are then used to estimate a probability density function for climate sensitivity by giving greater weight to smaller differences. Such inferences complement other approaches to the assessment of climate sensitivity based on varying uncertain parameters in climate models and evaluating the quality of the different model realizations based on their skill in reproducing observed *mean* climate (see Chapter 10, Section 10.5.4.4). While observed climate changes offer 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 (see 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.

Studies that base their assessments on the ability of climate models to simulate observed climate change vary key climate and forcing parameters (referring both to tuning parameters and system properties) 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. Parameters such as the climate sensitivity can be varied directly in simple climate models or in some earth system models of intermediate complexity (the so-called EMICS; see Chapter 8), and indirectly by varying model parameters in more complex EMICs and AOGCMs. Since studies estimating climate sensitivity from observed climate change require very large ensembles of simulations of historical climate change (ranging from several hundreds to thousands members), they are most 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 the likelihood of each parameter value given the data (see Kennedy and O'Hagan, 2001). The uniform distribution is one such prior distribution, which specifies no knowledge or assumptions about the parameters apart from the range of possible values. Uniform priors in equilibrium climate sensitivity have been used widely, which enables comparison of constraints obtained from the data in different approaches. 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 the equilibrium climate sensitivity, due to the nonlinear relationship between the two parameters. A similar problem arises with the transient climate sensitivity. Since the observational constraints on the climate sensitivity are still weak (as will be shown below), these prior assumptions influence the resulting estimates. Frame et al. advocate sampling the flat ("uninformative") prior in equilibrium sensitivity if this is the target of the estimate, or in transient climate response if the 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 et al. (2005) and Hegerl et al. (2006), multiple and independent lines of evidence about climate sensitivity from, for example, analysis of climate change at different times, can be combined, using information from one line of evidence as prior information for analysis of another line of evidence. In the following, uniform priors on the target of the estimate are used unless otherwise specified.

Section 9.6.2 discusses estimates of climate sensitivity from instrumental observations of long-term changes and specific, well observed events in recent climate history, such as the eruption of Mt. Pinatubo in 1991. This is followed by Section 9.6.3, which discusses estimates based on paleoclimatic data of the last millennium and the Last Glacial Maximum. A summary of methods, results and uncertainties is given in Table 9.6.1 (note that inverse aerosol estimates, which can be derived from some results, are also listed in Table 9.2.1). 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 true uncertainty of the 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 long-term changes in surface, ocean and atmospheric temperatures to estimate climate sensitivity. Most studies use the observed surface temperature changes over the 20th century or the last 150 years (see Chapter 3). In addition, some studies also use the estimated ocean heat uptake since 1955 based on data from Levitus et al. (2000; 2005) (see Chapter 5), and temperature changes in the free atmosphere (see Chapter 3; see also Table 9.6.1). For example, Frame et al. (2005) and Andronova and Schlesinger (2000) use surface air temperature alone, while Forest et al. (2002; 2006), 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 radiosonde data for atmospheric temperature change in addition to surface air temperature. Note that studies using radiosonde data may be affected by recently discovered inhomogeneities in radiosonde data (see 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 with it, equilibrium climate sensitivity) from observed changes in forcing and climate.

[INSERT TABLE 9.6.1 HERE]

Wigley et al. (1997) pointed out first that uncertainties in forcing and response made it impossible to use observed global temperature changes to constrain equilibrium climate sensitivity 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 climate sensitivity of 2.0°C and an upper limit for fossil fuel and biomass aerosol forcing (see Section 9.2.1.2) based on the range of 1.5 to 4.5°C for equilibrium climate sensitivity. 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. This hypothesis lacks plausibility since it is unclear why the surface should lag the troposphere. Further, their estimate does not account for internal climate variability or the role of external forcing. The finding of Lindzen and Giannitsis is also in contrast with that of Forest et al. (2002; 2006) who considered the joint evolution of surface and upper air temperatures on longer timesales.

Recently, a number of new results have become available that derive probability estimates for equilibrium climate sensitivity, which are described below. The range of models that have been applied varies widely. Gregory et al. (2002a) consider a simple box model of the change in the Earth's energy balance driven by the change in radiative forcing, the observed temperature change and the estimated change in ocean heat uptake. Other approaches use Energy Balance Models (EBMs, e.g. Andronova and Schlesinger, 2001; Frame et al., 2005) or EMICS (for example, Forest et al., 2002; for example, Knutti et al., 2002; 2003; 2006).
The diagnostics that compare model simulated to 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 optimal detection (see Section 9.4.1.4 and Appendix 9.A.1) based on zonal mean patterns of warming to diagnose the fit between model simulated and observed climate change patterns. Others do not directly apply multi-fingerprint detection and attribution studies. Frame et al. (2005) use detection results from based on several multi-model AOGCM fingerprints (see Section 9.4.1.4) that separate the greenhouse gas response from that to other anthropogenic and natural forcings (Stott et al., 2006c), and then compare the estimate of observed greenhouse gas warming with EBM simulations. Gregory et al. (2002a) apply an inverse estimate of the range of aerosol forcing based on fingerprint detection results. Note that results from fingerprint detection approaches will be affected by uncertainty in separation between greenhouse gas and aerosol forcing. However, as shown in Section 9.4.1.4, the resulting uncertainty in estimates of the near-surface temperature response to greenhouse gas forcing is relatively small (see discussion in Section 9.2.3).

A further important consideration in assessing these results is the use of realistic forcing estimates including forcing uncertainties. Most studies consider a wide 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 it using a wide range of direct forcing (for example, Andronova and Schlesinger, 2001; Forest et al., 2002; Forest et al., 2006). Many consider also tropospheric ozone (e.g., Andronova and Schlesinger, 2001; Knutti et al., 2002, 2003). Forest et al. (2006) demonstrate that the inclusion of natural forcing can considerably change probability density functions of climate sensitivity, since net negative natural forcing in the second half of the 20th century favors higher climate sensitivities than earlier results disregarding natural forcing (Forest et al., 2002), particularly if the same ocean warming estimates were used. Note however that recently revised ocean warming data (Levitus et al., 2005) favour somewhat smaller ocean heat uptakes than earlier data (Levitus et al., 2001) leading to less change in the probability density function due to inclusion of natural forcing (Forest et al., 2006). Almost all approaches account for aerosol forcing uncertainty, but only some studies account for natural forcing uncertainty or land surface changes, sometimes in sensitivity studies (Forest et al., 2006; see Table 9.2.1 for an overview). Some of the most thorough use of forcing uncertainty is in Knutti et al. (2002; 2003), who interpret the IPCC (2001)forcing uncertainty ranges as ±2 standard deviation confidence intervals around the best estimate.

The treatment of uncertainty in the ocean's uptake of heat also 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 9.6.1 compares results from many of these studies. Note that all plotted results use a uniform prior on equilibrium climate sensitivity and that they have been rescaled to integrate to unity for all positive sensitivities up to 10°C. Thus zero prior probability is assumed for sensitivities exceeding 10°C, and also for negative sensitivities (note that negative sensitivity would lead to an unstable climate and is thus inconsistent with the climate system). This figure shows that best estimates of the equilibrium climate sensitivity (mode of the estimated probability density functions, see Figure 9.6.1 and Table 9.6.1) typically range between 1.2 and 4°C when inferred from constraints provided by historical instrumental data, in agreement with other estimates derived from comprehensive climate models. Results from 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. Such studies cannot rule out, with reasonable likelihood, the possibility that the climate sensitivity substantially exceeds 4.5°C. For example, Andronova and Schlesinger (2001) find a greater than 50% likelihood that climate sensitivity lies outside the 1.5 to 4.5°C range. Knutti et al. (2002; 2003) and Gregory et al.(2002a) found no reliable upper limit on climate sensitivity based on analyses of global mean ocean and surface temperature changes. Forest et al. (2006) find a 5–95% range of 2.1 to 8.9°C for climate sensitivity (Table 9.6.1), 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 well beyond 10^{o}C sensitivity.

Knutti et al. (2002) also find that the available data on ocean heat uptake does not constrain their ocean mixing parameter, while Forest et al. (2002; 2006) find constraints on ocean diffusivity that suggest that models may be mixing heat too efficiently into the ocean, although this finding is uncertain due to structural model uncertainty. Knutti et al. (2002) also determine that strongly negative aerosol forcing, as has been suggested by several observational studies (see Anderson et al., 2003), is incompatible with the observed warming trend over the last century (see Section 9.2.1.2 and Table 9.2.1).

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 rather than uniform priors (see discussion above and in Appendix 9.B). Frame et al. (2005) find that sampling uniformly in transient climate response results in an estimated equilibrium climate sensitivity of 1.2 to 5.2°C with a median value of 2.3°C. Also, several approaches have been based on estimates of climate feedback and apply a uniform prior distribution on feedbacks. Translating these results into estimates of the equilibrium climate sensitivity is equivalent to using a prior distribution that favors smaller sensitivities, and hence tends to result in narrower ranges of equilibrium climate sensitivity (Frame et al., 2005). Forster and Gregory (2006) estimate climate sensitivity 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% equilibrium climate sensitivity range of 1.0 to 4.1°C (based on a prior that emphasizes lower sensitivities stronger as discussed above). The climate feedback parameter estimated from IPCC AR4 AOGCMs ranges from about 0.7 to 2.0 W/(m^2 K) (see Chapter 8, Table 8.8.1). Forster and Gregory (2006) tentatively find that their results may support neutral or negative longwave feedback rather than positive as in models.

[INSERT FIGURE 9.6.1 HERE]

9.6.2.2 Estimates based on individual volcanic eruptions

Some recent analyses have attempted to derive insights into the equilibrium climate sensitivity from the well observed forcing and response to the eruption of Mount Pinatubo, or other major eruptions during the 20th century. Such events allow for the study of physical mechanisms and feedbacks as was done by Forster and Collins (2004; see, Figure 8.6.2) or Soden et al. (2002) who demonstrated agreement between simulated and observed responses based on a model with a climate sensitivity of 3.0°C, unless the water vapour feedback in that model was 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 6.3°C, and determine that the cloud feedback in the latter model appears inconsistent with data. Note that this result may be specific to the model analyzed and hence cannot rule out a high sensitivity.

Constraining equilibrium climate sensitivity from the observed response to individual volcanic eruptions is difficult because the response to short-term volcanic forcing is strongly nonlinear in equilibrium climate sensitivity, yielding only slightly enhanced peak responses and substantially extended response times for very high sensitivities (Frame et al., 2005; Wigley et al., 2005a). The latter are difficult to distinguish from a noisy background climate. A further difficulty is uncertainty in the rate of heat taken up by the ocean in response to a short, strong forcing. Wigley et al. (2005a) 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 range of 1.5 to 4.5°C and find 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, and other eruptions suggest different upper limits. 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. However, Wigley et al. (2005b) demonstrate that the analysis method of Douglass and Knox (2005) severely underestimates climate sensitivity (by a factor of 3) if applied to the volcanic response in a climate model with known climate sensitivity, due to inadequate treatment of the effect of the ocean heat uptake. Furthermore, as pointed out by Frame et al. (2005), the effect of noise on the estimate of the climatic background level can substantially distort the estimate if not accounted for.

In summary, the responses to individual volcanic eruptions, particularly that to the well-observed eruption of Mount Pinatubo, provide a powerful test for feedbacks in climate models. However, due to the physics involved in the response, such individual events cannot provide tight constraints on equilibrium climate sensitivity. 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 the equilibrium climate sensitivity is the equilibrium global mean temperature change that eventually results from CO_2 doubling, the smaller *transient climate response* 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 Chapter 8, Section 8.6.2.1). The transient climate response is indicative of the temperature trend associated with external forcing, and so is constrained by an observable quantity, the observed warming trend that is attributable to greenhouse gas forcing. However, the transient climate response does not scale linearly with equilibrium climate sensitivity because transient sensitivity is strongly influenced by the speed with which the ocean transports heat into its interior, while equilibrium response is governed by feedback strengths (see discussion in Frame et al., 2005). It is therefore difficult to rule out high values of equilibrium climate sensitivity on the basis of a transient response. This is exacerbated by uncertainty in aerosol forcing (Boucher and Haywood, 2001; Chapter 2, Section 9.2), with the result that high values of equilibrium climate sensitivity can be partly offset by strong aerosol forcing during the 20th century. .

Transient climate response may be more relevant to determining near term climate change than equilibrium climate sensitivity. Stott et al. (2006c) estimate the transient climate response 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 9.6.2) 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 CO_2 doubling (based on a 1% increase in CO_2), with a 5–95% range of 1.5 to 2.8°C. Note that since transient climate response 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 the equilibrium climate sensitivity, although the separation of greenhouse gas response from other external forcing in a multi-fingerprint analysis introduces a small uncertainty, illustrated by small differences in results between 3 models (see Figure 9.6.2). Deflating significance levels in order to account for structural uncertainty in the estimate, it is unlikely that transient climate response is more than 2.8°C/century. The implications of this for future climate change are discussed in Chapter 10, Section 10.5.4.5.

[INSERT FIGURE 9.6.2 HERE]

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 state 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 of feedbacks in climate models.

As with analyses of the instrumental record discussed in Section 9.6.2, some studies using paleoclimatic data have also quanitified the agreement between past climate and simulations with model versions with varying sensitivity to CO_2 doubling in order to derive a probability density function for equilibrium climate sensitivity. Inferences about equilibrium climate sensitivity made through direct comparisons between radiative forcing and climate response, without using climate models, show large uncertainties. Problems are, for example, that climate feedbacks may be different for different climatic background states, and for different seasonal characteristics of forcing. Thus, the sensitivity to past climatic forcing may be substantially different from that to a doubling of CO_2 for past climate states with a strong seasonal change in insolation (see, for example, Montoya et al., 2000), and hence sensitivity to forcing cannot be directly compared to that to a doubling of CO_2 .

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 (see Chapter 6). To account for the uncertainty in reconstructions, Hegerl et al. (2006) 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 and a new reconstruction) to constrain estimates of the climate sensitivity during the pre-industrial period up to 1850. As with the studies discussed in Section 9.6.2, a large ensemble of simulations of the last millennium is performed with an energy balance model that is forced with reconstructions of volcanic (Crowley, 2000, updated), solar (Lean et al., 2002), and greenhouse gas forcing (see Section 9.3.3 for results on the detection of these external influences). The estimated probability distribution functions for equilibrium climate sensitivity incorporate uncertainty in the overall amplitude, but not the shape, of volcanic and solar forcing. Uncertainty in the amplitude of the reconstructed temperatures associated with calibration is accounted for in one reconstruction, and by a sensitivity study in all others. Probability density functions for climate sensitivity from all reconstructions combined yield a median sensitivity of 3.4°C and a 5–95% range of 1.2 to 8.6°C, and a shape and tail of the pdf that is quite similar to those from the instrumental period, (see Figure 9.6.1). Reconstructions with a higher amplitude of past climate variations (such as, Esper et al., 2002 or the new reconstruction) are found to support higher climate sensitivity estimates than reconstructions with lower amplitude (such as, Mann and Jones, 2003).Note that the constraint on climate sensitivity originates mainly from low-frequency temperature variations associated with changes in the frequency and intensity of volcanism that leads to a highly significant detection of volcanic response (see 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 Northern Hemispheric mean temperature from 1500 onward when compared to the Mann and Jones (Mann and Jones, 2003) reconstruction. The climate sensitivity was determined by the best match between a simple climate model with varying climate sensitivity, driven by a variety of estimates of radiative forcing, and instrumental data from 1861 to the present.

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 (see Section 9.2.1.3). Different Northern Hemisphere temperature reconstructions (see 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 estimates of climate sensitivity based on a comparison between radiative forcing and climate response is subject to large uncertainties. Best guess estimates depend on the reconstruction used, but are broadly similar to estimates discussed above. Again, reconstructions with stronger cooling in this period support higher climate sensitivities than those with smaller amplitude (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 is one of the key periods used to estimate constraints on climate sensitivity (see the pioneering work of Hansen and al., 1984; see also studies such as Lorius et al., 1990; Hoffert and Covey, 1992).The most recent AOGCMs run for the climate of the Last Glacial Maximum show that changes in greenhouse gas concentrations and in the extent and height of ice sheet boundary conditions produce cooling estimates from 3.5–5.2°C (see Chapter 6, Section 6.4.1 and Section 9.3.3, see also Masson-Delmotte et al., 2005). These model responses are in broad agreement with the data, suggesting that they adequately represent the feedback mechanisms that determine the climate sensitivity that corresponds to the LGM climate state (note that the models were not retuned for the purpose of simulating the LGM (see Section 9.3.3). In these models, the global surface cooling is the response of the climate system to radiative perturbations of 4.1–7.2 W m^{-2}, and the equilibrium climate sensitivity of the models used in PMIP2 ranges from 2.3 to 3.7°C for CO_2 doubling. Note, however, that this range of simulated climate change may not reflect the full uncertainty as it also may be affected by incomplete forcings in the PMIP2 LGM experiment and large uncertainties in forcing and data (see Chapter 6 and also Section 9.2.1.3). In addition, the range of simulated temperature changes is affected by differences of the radiative influence of the ice-covered regions in different models (Taylor et al., 2000).

Two studies (Annan et al., 2005; Schneider von Deimling et al., 2006) have now attempted to estimate the probability density function of equilibrium climate sensitivity from ensemble simulations of the Last Glacial Maximum 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, to infer bounds on climate sensitivity. Annan et al. (2005) used a low-resolution version of the CCSR/NIES/FRCGC atmospheric GCM coupled to a mixed layer ocean. An ensemble Kalman filter method was used to obtain a posterior joint probability distribution for 25 uncertain parameters when tuned to reproduce observed seasonal climatological fields of 15 variables, and 40 member ensemble simulations were made of the response to both doubled CO_2 and Last Glacial Maximum (LGM) forcings. They find a best-fit sensitivity of about 4.5°C, and results that suggest that sensitivities in excess of 6°C are unlikely given observed estimates of LGM tropical cooling and the relationship between tropical SST and sensitivity in their model. In another perturbed perturbed physics ensemble, Schneider von Deimling et al. (2006) vary 11 ocean and atmospheric parameters in 1,000 member ensemble simulations of the CLIMBER-2 EMIC (see Chapter 8, Table 8.8.2). They found a different relationship between sensitivity and tropical SST cooling, implying a low probability for sensitivities exceeding 3.5°C based on their model only, and a 5–95% range of 1.2 to 4.3°C when attempting to account more fully for uncertainties including structural uncertainties. Schneider von Deimling et al. (2006) find similar constraints on climate sensitivity if proxy reconstructions of LGM Antarctic temperatures are used instead of tropical SSTs. The discrepancy between the inferred upper limits in the two studies arises from structural differences between the models used: the model used by Annan et al. (2005) shows a much weaker connection between simulated tropical SST changes and climate sensitivity than that used by Schneider von Deimling et al. (2006). This illustrates that it is important to sample uncertainties within different model frameworks to establish the robustness of simulated relationships between observables and future changes. A further structural uncertainty is that both studies use simplified ocean models, which may not capture the full ocean response affecting tropical SSTs. Thus these results broadly support other estimates of climate sensitivity derived, for example, from the instrumental period, but they are presently unable to constrain them further.

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 (see 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 (climate sensitivity, radiative forcing, mixing of heat into the ocean) have been taken into account in most studies (Table 9.6.1), some caveats remain. Some processes and feedbacks might be poorly represented or missing, particularly, but not only, in simple or many intermediate complexity models. Thus, structural uncertainties in the models that simulate changes with different climate sensitivities (such as problems in representation of clouds feedback processes, see Chapter 8, and of the physics of ocean mixing, or uncertainties in the connection between tropical SST changes and sensitivity in models) will affect results and are very difficult to quantify. The use of a single value for equilibrium climate sensitivity in terms of 'effective' climate sensitivity (see Glossary) further assumes that the climate sensitivity is constant in time. However some authors (e.g., Senior and Mitchell, 2000; Boer and Yu, 2003) have shown that the effective climate sensitivity 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. Furthermore, there is a small uncertainty in the radiative forcing due to CO_2 doubling (<10%; see Chapter 2), which is also not accounted for in most studies deriving observational constraints.

Despite these uncertainties, which are accounted for to differing degrees in the various studies, confidence is increased by similarities between individual estimates of equilibrium climate sensitivity (Figure 9.6.1). Most studies find a lower 5% limit of 1°C or greater, and a most likely value of climate sensitivity between 1 and 4°C, centred around 2–3°C. 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 equilibrium climate change 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, in part, 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. 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. For the last millennium, uncertainties in temperature and forcing reconstructions, and the nonlinear connection between sensitivity and the response to volcanism prohibit tighter constraints. A tight constraint based on the sensitivity of the Last Glacial Maximum climate is difficult to obtain because of the present uncertainty regarding tropical temperature changes. In addition, differences in the results of Annan et al. (2005) and Schneider von Deimling et al. (2006) suggest that tropical SSTs cannot yet be reliably used to estimate global equilibrium climate sensitivity given the effect of structural uncertainties on the connection between CO_2 forcing and tropical SSTs in these models.

Thus most studies that use a simple uniform prior on the sensitivity are not able to exclude equilibrium climate sensitivity beyond the traditional 1.5 to 4.5°C range although the combined body of work does now help to quantify the likelihood that it lies in this range (see Section 9.6.4). However, 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.

Bayesian methods allow for incorporating multiple lines of evidence to sharpen the posterior distribution of climate sensitivity. Such an approach has been taken in two recent studies (Annan and Hargreaves, 2006; Hegerl et al., 2006). Annan et al. (2006) demonstrate using three lines of evidence, namely 20th century warming, response to individual volcanic eruptions, and LGM response, results in tighter estimate of the equilibrium climate sensitivity, with a probability of less than 5% that equilibrium climate sensitivity exceeds 4.5°C. A similar constraint arises if 5 lines of evidence are used 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 Chapter 10, Box 10.2). However, as discussed in Annan et al. (2006),such a combination of lines of evidence is reliable only if every single line of evidence is entirely independent of others, an assumption that is easily violated if, for example, individual volcanic eruptions also influence the long-term climate trend or due to other overlaps. Hegerl et al. (2006) argue that instrumental temperature change during the second half of the 20th century is essentially independent of the paleo record and of the instrumental data from the first half of the 20th century that is used to calibrate the paleo records. Hegerl et al. (2006) therefore base their prior probability distribution for the climate sensitivity on results from the late 20th century (Frame et al., 2005), which when updated reduces the 5–95% range of equilibrium climate sensitivity 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.

Thus, several lines of evidence strengthen confidence in present estimates of equilibrium climate sensitivity, leading to the overall conclusion that climate sensitivity is likely to lie in the range of 2 to 4.5°C (see Chapter 10, Box 10.2). Although tighter limits can theoretically be obtained by combining multiple lines of evidence, remaining uncertainties that are not accounted for in individual estimates (such as structural model uncertainties) and small dependencies between individual lines of evidence make a tighter constraint uncertain at present. The results from individual studies and the consistency of estimates from different time periods lead to the conclusion that it is very likely that climate sensitivity exceeds 1.5°C. This lower bound is consistent with the view that the sum of all atmospheric feedbacks affecting climate sensitivity is very likely positive. Constraints based on observed climate change support the assessment (see Chapter 10, Box 10.2) that equilibrium climate sensitivity is likely in the range of 2.0 to 4.5°C with a most likely value around 3.0°C. Values substantially higher than 4.5°C cannot be excluded, but agreement with observations and proxy data is generally worse for those high values than for values in the 2–4.5°C range.

9.7 Combining Evidence of Anthropogenic Climate change

The wide-spread change detected in temperature observations of the surface (see Sections 9.4.1-3), free atmosphere (see Section 9.4.4), and ocean (see Section 9.5.1), together with consistent evidence of change in other parts of the climate system (see Section 9.5), strengthens the overall evidence that greenhouse gas forcing is the dominant cause of warming during the past several decades. This combined evidence, which is summarized in Table 9.7.1, is substantially stronger than the evidence that is available from observed changes in global surface temperature alone (see Chapter 3, Figure 3.2.6).

[INSERT TABLE 9.7.1 HERE]

The evidence from surface temperature observations is strong: Climate models only reproduce the observed global mean surface warming when both anthropogenic and natural forcings are included (Figure 9.4.1). 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. Moreover, the response to anthropogenic forcing is detectable on all inhabited continents individually, and in many sub-continental regions. It is highly unlikely (<5%) that recent global warming is due to internal variability; such as might arise from El Niño (see Section 9.4.1). Paleo climatic evidence indicates that El Niño variability during the 20th century is not unusual relative to earlier periods (see Section 9.3.3.2 and Chapter 6). Moreover, a sustained warming from this mechanism would not be possible without a sustained source of heat. Moreover, the widespread nature of the warming (see Chapter 3, Figure 3.2.9 and Figure 9.4.2) reduces the possibility that the warming could have resulted from internal variability. Although modes of internal variability such as El Niño can lead to global average warming for limited periods of time, the warming is regionally variable, with some areas of cooling (see Chapter 3, Figures 3.6.2 and 3.6.3). No known mode of internal variability leads to such widespread, near universal warming as has been observed in the past few decades. 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 unlikely to be due to natural internal variability. Detection and attribution of external influences on these records, from both natural and anthropogenic sources (Figure 9.3.1, Table 9.7.1) 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 other components of the climate system such as the ocean or the cryosphere. 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, over the past half-century, the extent of all components of the cryosphere (glaciers, small ice caps, ice sheets, and sea-ice) has reduced, consistent with anthropogenic forcing (Section 9.5.4, Table 9.7.1), 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) has also increased, much more substantially than have the other components of the climate system (see Chapter 5, Table 5.2.1, Levitus et al., 2005, Hansen et al., 2005). The penetration of heat into the ocean is consistent with greenhouse gas forcing, and 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 9.7.1). Heating of the ocean resulting from net positive forcing explains the pattern of ocean heat content changes with warming proceeding from the upper layers of the ocean; deeper penetration of heat at middle to high latitudes and shallower penetration at low latitudes is consistent with our understanding of ocean structure. Thus the ocean was not the source of the warming at the surface and near surface. Simulations forced with observed SST changes cannot explain the the warming in free atmosphere without increases in greenhouse gases (Sexton et al., 2001), further strengthening the evidence that the warming does not originate from the ocean. The simultaneous increase in energy content of all the major components of the climate system indicates that the cause of the warming is highly unlikely to be the result of internal processes.

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, net climate forcings and current planetary imbalance in the radiation budget (Hansen et al, 2005). For the observed increase in ocean heat storage to have resulted from internal variability, surface ocean temperatures would have needed to cool to decrease outgoing heat flux, whereas the surface oceans have warmed. Only by a net positive forcing of the climate system, consistent with observational and model estimates of the likely net forcing of the climate system (as used in Figure 9.4.1), is it possible to cause the large increase in heat content of the climate system that has been observed (see Chapter 5, Table 5.2.1).

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Question 9.1: Can Individual Extreme Events be explained by Climate Change?

Several factors usually need to combine to produce an extreme event, so attributing a specific, single extreme event to a specific cause is difficult, if not impossible. As well, a wide range of extreme weather events is feasible even in an unchanging climate. However, the human-induced warming over the past century suggests that the likelihood of heat waves could have increased over the 20th century and that the likelihood of frost could have decreased. For example, a recent study suggests that human influences may have more than doubled the risk of a European summer as warm as that of 2003. Quantifying the influence of climate change on other types of individual extreme events such as floods or hurricanes may be more difficult because of greater natural variability and greater uncertainties in predicting the effect of climate change on such events.

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 examples of extreme weather that some commentators have linked to climate change. These include the prolonged drought in Australia, the extremely hot summer in Europe in 2003 (see Question 9.1, 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 atmospheric greenhouse gas concentrations have "caused" any of these events?

[INSERT QUESTION 9.1, FIGURE 1 HERE]

A wide range of extreme weather events is possible even with an unchanging climate, so it would be difficult to attribute an individual event, by itself, to a changed climate. As well, extreme weather results from a combination of factors. For example, dry soil (which leaves more solar energy available to heat the land because less energy is consumed evaporating moisture from the soil), a persistent blocking high pressure system, and very clear skies, all contributed to the very warm European summer of 2003. 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 influences (e.g., sea surface temperatures) but others may not, this will complicate the detection of a human influence on a single, specific extreme event.

However, we may be able to determine whether anthropogenic forcing has changed the likelihood of occurrence of a specific type of extreme weather event such as heat waves. This can be addressed, for example for the 2003 European heat wave, by studying the characteristics of European summers in a climate model, either forced only with historical changes in natural factors such as volcanic activity and the solar output, or by both human and natural factors. Such experiments indicate that over the 20th century, human influences may have more than doubled the risk of European mean summer temperatures as hot as those in 2003. 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 a probability-based approach ("is there a change in the likelihood of an event that results from human influence?") is that it can be used to estimate the influence of external factors, such as increases in greenhouse gas concentrations, on the frequency of specific types of weather events (e.g., frost). However, careful statistical analyses are required, since the likelihood of individual extremes, such as a late-spring frost, could change due to changes in variability as well as changes in the mean climate. Such analyses rely on climate model-based estimates of variability, and thus an important additional requirement is that climate models adequately represent climate variability.

The same likelihood-based approach could be adopted to examine possible changes in the frequency of extreme hydrological events such as heavy rainfalls or floods. Climate models predict that there will be changes in the incidence of many types of extreme weather events, including an increase in extreme rainfall events, due to human influences on the atmosphere. There is some evidence of increases in extreme rainfall events in at least some regions in recent decades. However there is as yet no conclusive evidence that these increases are necessarily linked to increasing greenhouse gas concentrations in the atmosphere.

Question 9.2: Can the Warming of the 20th Century be explained by Natural Variability?

The warming over the last century has been very unusual in both magnitude and rate of change relative to other periods in the last millennium according to reconstructions of temperature from paleoclimatic records such as tree rings and ice cores. This rapid warming is consistent with physical understanding of how the climate should respond to the observed rapid increase in greenhouse gas concentrations, and inconsistent with the response expected from natural external influences on the climate system (e.g., variability in solar output and volcanic activity). Climate models provide good simulations of large-scale temperature variations of the last 100 years when they include the dominant external factors believed to have affected the climate system during this period but fail to reproduce the recent warming when human-induced forcings (e.g., increased greenhouse gases) are not included. Spatial patterns of temperature change observed in recent decades are similar to those expected from increased greenhouse gas concentrations and differ in many respects from the most important patterns of natural climate variability such as the El Niño – Southern Oscillation (ENSO). Climate models almost never simulate warming rates of the magnitude observed over the 20th century when external influences (natural and human-induced) on the climate system are held constant, even in very long multi-millennium simulations. Based on these various lines of evidence it is very unlikely that the 20th century warming can be explained by natural variability, either from internal sources (e.g., ENSO) or natural external influences (e.g., change in solar output or volcanic activity).

The climate varies over time because of natural internal processes such as ENSO and also because of changes in external influences (forcings). These external influences can be both natural in origin (arising principally from aerosols emitted by volcanic eruptions, and from variations in solar output) and human-induced (arising from factors such as greenhouse gas emissions, human-sourced aerosols, ozone depletion, and land use change). The role of natural internal processes can be estimated by running climate models without changes in external forcings. The effect of external forcings, including the geographical patterns of change expected from the various forcings, can be estimated with models by changing the forcings, and by using physical understanding of the processes involved. The combined effects of natural internal variability and natural external forcings can also be estimated from paleoclimatic reconstructions of pre-industrial climate variations.

The multi-decadal temperature variations observed over the 20th century are consistent with scientific understanding of how the climate is expected to respond to changes in both human-induced and natural external forcing. Natural external forcing factors include emissions of volcanic aerosols and variations in solar output. Explosive volcanic eruptions occasionally eject large amounts of dust and sulphate aerosol high up in the atmosphere temporarily shielding the earth and reflecting sunlight back to space. Solar output varies on an 11 year cycle and could also have longer term variations. Over the last 100 years, there has been a rapid increase in the atmospheric content of carbon dioxide and other well mixed greenhouse gases that had previously remained at near stable concentrations in the atmosphere for thousands of years. Human activities also led to increased concentrations of aerosols of sulphur and other chemicals in the atmosphere, particularly in the 1950s and 1960s.

Although internal variations in the climate system such as ENSO can cause variations in global temperature, a large fraction of the multi-decadal change in global mean temperature over the 20th century is likely to have been caused by external forcings. Short-lived decreases in global temperature have followed volcanic eruptions such as Pinatubo in 1991. There was global mean warming in the early part of the 20th century, a period during which greenhouse gas concentrations had started to rise and before tall chimney stacks increased atmospheric lifetimes of sulphate aerosols, volcanic activity was reduced, and solar output was likely increasing. The levelling off of global mean temperatures in the 1950s and 1960s accompanied increases in human-sourced aerosols that tend to cool the planet by reflecting radiation directly to space and by making clouds more reflective and longer lasting. The eruption of Mt. Agung in 1963 also put large quantities of reflecting dust in the upper atmosphere. The warming observed since the 1970s has occurred in a period when greenhouse gas forcing has dominated over all other forcings.

The warming patterns observed in the most recent four decades are consistent with physical understanding of greenhouse gas forcing. There has been a warming of the lower atmosphere (the troposphere) and cooling higher up in the stratosphere. Also, at the surface there has been a greater warming over land than ocean with the largest warming at high northern latitudes. Such patterns of change are not only consistent with those expected from human-induced forcing but they also differ in many respects from the principal patterns of temperature change associated with internal variability, such as those related to ENSO.

Numerous experiments have been conducted with coupled ocean-atmosphere climate models, forced with natural and/or human-induced forcings, to determine the likely causes of the 20th century climate change. These experiments, conducted with a variety of models, indicate that models cannot reproduce the rapid warming of the surface observed in recent decades when they use only natural external forcing (i.e., when they only take into account variations in solar output and volcanic aerosols). However, as shown in Question 9.2, Figure 1, models are able to simulate the observed 20th century changes in temperature when they include all of the most important forcings, including forcings from human sources (changes in greenhouse gases, stratospheric ozone, and industrial aerosols) as well as natural external forcings. Model simulated responses to external forcing are detectable in the 20th century climate both globally and in each individual continent except Antarctica. The effect of human-induced forcing very likely dominates over all other causes of global mean surface temperature change during the past half century. Long multi-century climate simulations have also been used to estimate the variability of the climate system in the absence of natural and human-induced external forcing. Century scale temperature changes that result from internal climate variability in such simulations are quite small relative to the strong 20th century warming, suggesting that internal climate system mechanisms do not provide an explanation for the observed warming.

[INSERT QUESTION 9.2, FIGURE 1 HERE]

An important source of uncertainty arises from the incomplete knowledge about some of forcings, such human-sourced aerosols, used in model experiments. In addition, the climate models themselves are imperfect. However models exhibit a common pattern of response to human-induced greenhouse forcing that includes greater warming over land than ocean, warming in the troposphere and cooling in the stratosphere. Such characteristic patterns help to distinguish the response to greenhouse gases from that to natural external forcings, for example from changing solar output which would have a different pattern of response in the troposphere and stratosphere. In addition, the different time histories of the human-induced and natural external forcings help to distinguish between the responses to these forcings. Such considerations increase confidence that the global warming observed over the last 50 years was dominated by human-induced rather than natural external forcings.

Reconstructions of Northern Hemispheric temperature of the last 1-2 millennia, based on “proxy” data (e.g., tree rings that vary in width or density from one year to the next in step with temperature changes) provide another line of evidence that natural internal variability and natural external forcings cannot explain the 20th century warming. While there is some uncertainty in the estimates of past temperature variability, all reconstructions indicate that the warming observed during the 20th century is stronger than any other century-scale global temperature change during at least the last 1000 years. Much of the variation in the reconstructed record of Northern Hemisphere mean temperature prior to the industrial era can be explained by variations in natural (volcanic and solar) forcing. However, the natural forcings have been much smaller than those from human sources over the last 100 years. An adequate explanation of the temperature record during the industrial era is only possible if both human-induced forcings and natural forcings are taken into account.

Tables

Table 9.2.1. Inverse estimates of aerosol forcing from detection and attribution studies and studies estimating equilibrium climate sensitivity (see Table 9.6.1 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*: 5–95% inverse estimate of the total aerosol forcing in the year given relative to preindustrial forcing. Note that 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 or not explicitly considered forcings (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: fossil fuel and biomass burning black carbon and organic matter. Values are taken from publications and depend on ranges considered, priors, etc. (see Table 9.6.1). Note that for results where OzT and Fossil fuel and biomass burning black carbon and organic matter is not directly accounted for, the forcing by them will be incorporated in the total aerosol forcing estimated.

Study	*Forest et al. (2002)*	*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 Temp., latter half 20th century	Same as 2002, updated ocean temperatures	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, OzS	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
Aerosol forcing [W/m^2]	–0.3 to –0.95 –0.25 to –0.9 with expert prior 1980ies	–0.14 to –0.74 –0.07 to –0.65 with expert prior 1980ies	–0.54 to –1.3 1990	0 to –1.2 ind. aerosol, –0.6 to –1.7 total aerosol 2000	–0.4 to –1.6 total aerosol by 2000	–0.4 to –1.4 total aerosol by 2000	Fossil fuel aerosol unlikely <–1, biomass+dust unlikely <–0.5[a] (1990)

Notes:
(a) Explores IPCC range of climate sensitivity, while other studies explore wider ranges

Table 9.3.1. Detection and attribution results for a range of paleoclimatic records of the last millennium based on a multiple regression of records onto the simulated response to solar, volcanic, and greenhouse gas and aerosol (ghg+aer) forcing. “Y” or “N” indicates that the response to external forcing is, or is not, detectable. The bottom row gives the standard deviation of the (decadal) residual (in ^{o}C). The values in parentheses in that row indicate the percentage of decadal variance in each proxy reconstruction that is explained by external forcing. Updated after Hegerl et al. (2003). The “?” for greenhouse gases and aerosol signals in the Moberg et al. record refers to the results being sensitive to details of the analysis.

Record Period	Briffa et al. (2001) 1400–1940	CH-blend (Hegerl et al., 2006) 1270–1960	Mann et al. (1999) 1400–1960	Esper et al. (2002) 1270–1960	(Crowley and Lowery, 2000) 1000–1960	(Moberg et al., 2005) 1270–1925
Volcanic	Y	Y	Y	Y	Y	Y
Solar	N,	N,	N (Y for periods)	N	N (Y from 1100 on)	Y
Ghg + aer	Y	Y	N (Y to 1980)	Y	Y	?
Residual std	0.09 (57%)	0.09 (70%)	0.07 (49%)	0.15 (60%)	0.10 (57%)	0.11 (61%)

Table 9.6.1. Results from key studies on observational estimates of the equilibrium climate sensitivity parameter α (in °C) and the total or net fossil fuel-related aerosol forcing "aer" (in W m^{-2}) from instrumental data (white background), data for the last millennium (yellow), individual volcanic eruptions (purple) and simulations of the Last Glacial Maximum (blue). The first column lists the study, the second the observational data used to constrain sensitivity and aerosol forcing, the third the type of model used, the fourth the external forcings accounted for in the study, and the fifth the free parameters varied in the study (such as ocean diffusivity κ , or total aerosol forcing) . The sixth column gives the resulting estimate of equilibrium climate sensitivity, and the seventh of the most likely value found (mode, which is the peak of the probability pdf). Ideally, studies account for model uncertainty (in simple models, that due to uncertainty in ocean diffusivity κ), 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"). The uncertainties accounted for in individual studies (apart from the parameters varied) are listed in the last column. Values are taken from publications and depend on ranges considered, priors etc.
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 8.8.2)

Study	Data	Model	Forcing	Free Parameters	α Range 5–95% [°C]	Mode	Treatment of Incorporated Uncertainties
Forest et al. (2002)	Upper air, surface and deep ocean space-time 20th C Temp. (diag 2nd half)	2-D EMIC (~E6, Chapter 8, Table 8.8.2)	G, Sul, OzS,	α (0–10), κ , aer (0 to –1.5)	1.4 to7.7 (1.3–4.2 with expert prior)	2.0	ε_{obs} , noise (AOGCM)
Forest et al. (2006)	"	"	G, Sul, Sol, Vol, OzS, land surface changes	"	2.1 to 8.9 (1.9-4.7 with expert prior)	3.0	ε_{obs} , noise (OAGCMs), tests for effect of nat. forcing unc.
Andronova and Schlesinger (2001)	Global mean and hemispheric difference in SAT 1856–1997	EBM with ocean	G (detailed), OzT, Sul, Sol, VolSuli	α , choice of radiative forcing factors	1.0 to 9.3 p>54% that α lies outside 1.5–4.5	1.2	Noise (bootstrap residual)
Knutti et al. (2002; 2003)	Global mean ocean heat uptake 1955–1995, mean SAT inc. 1860–2000	EMIC ~E1, (+ neural net)	G, OzT, OzS, fossil fuel and biomass burning BC+OM, strat water vapour, Vol, Sol	α (1–10), κ , scaling for all forcings including aer	2.2 to 9.2 p> 50% that α lies outside 1.5–4.5	Median 4.9	ε_{obs} ε_{forc} from IPCC (2001), different ocean mixing schemes
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, 2001 #717}, Sol, Vol	Directly estimated	1.1 to inf with BC, 1.6 to inf. without BC	2.1	ε_{obs} , ε_{forc}
Frame et al. (2005)	Global change in surface temperature	EBM	G, accounted for other anthrop and natural forcing, by fingerprints, Sul, Nat?	Range of κ consistent with ocean warming	1.2 to 11.8 (sampled to 20) (0.4 to 4.0 if unif. feedbacks)	2.0	Noise, forcing uncertainties

Forster and Gregory (2006)	1985–1996 ERBE data 60N–60S, global surface T	1-Box	Ghg, Vol, Sol, Sul	Directly estimated	1.0 to 4.1 sampled unif. in feebacks	NA	ε_{obs} , forcing uncertainty discussed
Wigley et al. (2005a)	Global mean surface temperature	Simple model based on fit of exponential decay and max coling	From volcanic forcing only	estimated indirectly via decay time	Agung: 1.3–6.3 El Chichon: 0.3–7.7 Pinatubo: 1.8–5.2	Mean: Agung 2.8 El Chichon 1.5 Pinatubo 3.0	El Nino (missing: volcanic forcing uncertainties, noise other than El Nino)
Hegerl et al. (2006)	NH mean SAT preindustrial (1270/1505 to 1850) from multiple reconstructions	EBM	G, Sul, Sol, Vol	α (0–10), κ , aer (–0.3 to 2.5)	1.1 to 8.6 for rec. combined, 1.4–6.1 for CH rec.,	2.0 all com., 2.3 CH-b	ε_{obs} , ε_{nat} , ε_{aer} , noise (from residual)
Schneider von Deimling et al. (2006)	LGM SSTs and other LGM data	EMIC (~E3)	LGM forcing: greenhouse gases, dust, ice sheet b.c.s, vegetation, insulation	Full equilibrium, NA	1.2 to 4.3°C	Not given	Obs. uncertainty of ice age SSTs (one type of data); attempt to account for structural unc., some forcing unc,
Annan et al. (2005)	Present day seasonal cycle of a number of variables for sampling prior, validation through LGM SSTs	AGCM with mixed layer ocean	PMIP2 LGM forcing	“	<7% chance of sensitivity >6°C	4.5°C	Obs. uncertainty in tropical SST estimates (one type of data)

Table 9.7.1. A synthesis of climate change detection results. **a)** Surface and atmospheric temperature evidence. 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. **b)** Evidence from other variables. The likelihood assessment is indicated in parentheses where the level differs from one of the standard IPCC 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	Highly 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 (<10%). 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. Main uncertainties arise from forcing, including solar, model simulated responses and internal variability estimates (Sections 9.4.1.2, 9.4.1.4)
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. Main uncertainty from forcing and internal variability estimates (Section 9.4.1.4) .
Greenhouse gases would have caused more warming than observed over the last 50 years, with some warming offset by net cooling caused by natural and other anthropogenic factors.	Global	Likely	Estimates from different analyses using different models shows 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)
Anthropogenic warming during the past half century is detectable on every inhabited continent.	Africa, Asia Australia, Europe, North America and South America	Likely	Anthropogenic change has been detected at the 5% significance level on every continent and detectable anthropogenic warming is assessed as very likely on every continental region except Europe where greater variability compared to other continental regions makes detection more marignal (as illustrated by FAQ Figure 9.2.1). Uncertainties arise because sampling effects result in lower signal to noise ratio on smaller continental and subcontinental scales than on global scales. Separation of the response to different forcings is more difficult on these spatial scales (Section 9.4.2)

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. In the early part of the century, natural forcing is likely more important relative to anthropogenic forcing than later in the century, although there is likely a substantial warming contribution from anthropogenic forcing. Both natural forcing and response are uncertain. 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).
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. Uncertainty in reconstructions and past forcings; both uncertainties would tend to decrease similarity between simulation and reconstruction and hence reduce consistency between reconstructions and simulated response to past forcings. (Section 9.3.4)
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 increased temperature of warmest night, coldest day and coldest night annually has been formally detected in one study, but observed change in the warmest day is inconsistent with model result. This result 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 resulted mainly from greenhouse gas increases and stratospheric ozone changes. (Section 9.4.4.2). 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. Possibility of models having compensating errors in tropospheric warming and stratospheric cooling although overall tropopause height increases in recent model and observational datasets are driven by similar large scale changes in atmospheric temperature.
Tropospheric warming detectable and attributable to anthropogenic forcing (latter half 20th century)	Global	Likely	Robust detection 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 which is not seen in the radiosonde record (possibly due to uncertainties in the radiosonde record), but is seen in one processing of the satellite record, although not others. (Section 9.4.4).
Simultaneous tropospheric warming and	Global	Very Likely	Simultaneous warming of the troposphere and cooling of the stratosphere due to

stratospheric cooling due to anthropogenic forcing (latter half of 20th century)			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.

b)

Result	**Region**	**Likelihood**	**Factors contributing to likelihood assessment**
Ocean changes			
Anthropogenic forcing has warmed the upper ocean during latter half 20th century	Global (but with limited sampling in some regions)	Likely	Robust detection of anthropogenic fingerprint of ocean temperature changes in three different models suggests high likelihood, but observational and modelling uncertainty remains. 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)
Anthropogenic forcing is the largest contributor to sea level rise during the latter half 20th century	Global	Likely	Natural factors alone do not explain either the observed thermal expansion of the ocean or the observed sea level rise. Models including anthropogenic and natural forcing simulate the observed thermal expansion and rateof sea-level rise since 1961 reasonably well. Anthropogenic forcing dominates the change simulated by models. Models may have underestimated recent (1993–2003) sea level rise when contributions from all sources are considered. Modelled and observationally based estimates of variability of themal expansion component differ. Glacier mass balance sensitivity uncertain. Recent accelerations in ice flow in ice sheets is an effect that is not included in models.
Circulation			
Sea level pressure shows detectable anthropogenic signature during latter half 20th century	Global	Likely	Changes of similar nature observed in both hemispheres and 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 responses in the SH to ozone concentration changes are consistent with observations (Section 9.5.2.4).
Anthropogenic forcing contributed to increase in tropical cyclone intensity since the 1970s	Tropical regions	More likely than not (>50%) with low confidence	Recent observational evidence suggests an increase in frequency of intense storms. Increase in intensity 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. (Chapter 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	Model response detectable in observations for some models and result supported by

		(>50%)	theoretical understanding. However, uncertainties in models, forcings and observations. (Section 9.5.3.2). Limited observational sampling, particularly in the Southern Hemisphere.
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 uncertainty. (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, might 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)
Increase in continental runoff due to anthropogenic forcing during latter half 20th century	Global land areas	More likely than not (>50%)	Observed runoff increases, consistent with expectations. More than one mechanism may be responsible. Response to suppression of transpiration due to CO_2 -induced stomatal closure detected in one study. Large model and forcing uncertainties. (Section 9.5.3.2). Models capture some of the observed patterns of increases and decreases in runoff with some differences in magnitude of trends.
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 se 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. (Section 9.5.4.1). Observed and simulated changes not in agreement in the SH in one model.
Glacier retreat during 20th century due to anthropogenic forcing	Global	Likely	Formally detected with high confidence in one study (relative to internal variability) for only two glaciers, but observed changes qualitatively consistent with theoretical expectations and temperature detection. Volume change difficult to estimate. 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 frequently 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 inference 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 $\mathbf{a} = (X^T C^{-1} X)^{-1} X^T C^{-1} y = (\tilde{X}^T \tilde{X})^{-1} \tilde{X}^T \tilde{y}$ where matrix $\tilde{X}$ and vector $\tilde{y}$ represent the signal patterns and observations after normalization by the climate's internal variability. The normalization transform is performed to maximize the signal-to-noise ratio (see e.g., Mitchell et al., 2001).

The matrix $\mathbf{X}$ typically contains signals that are estimated with either a CGCM, 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 CGCMs simulate natural internal variability as well as the response to specified anomalous external forcing, the GCM 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). 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 simluated by CGCMs (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. CGCMs 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 $\mathbf{a}$. Most studies evaluate these hypotheses using standard frequentist methods (Hasselmann, 1979, 1997; Hegerl et al., 1997; Allen and Tett, 1999; Allen et al., 2005). 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, 2006).

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, 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).

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 (2006). 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, although their approach could be extended to also include the latter within the Bayesian framework.

Appendix 9.B: Methods Used to Estimate Climate Sensitivity and Aerosol Forcing

Two approaches have been used recently to estimate climate sensitivity, and other parameters such as aerosol forcing and ocean diffusivity, from observations. Gregory et al. (2002a) use a direct estimate of sensitivity based on uncertainties in components that determine climate response. Here, we focus on an approach that is closely related to climate change detection methods.

In this approach, observed climate change $T_{obs}(x,t)$, where *x* and *t* indicate space and time coordinates, is repeatedly compared to each of a series of climate change simulations $T(x,t,\theta)$ obtained from a climate model by varying the elements of a small vector θ of model parameters. A relatively simple climate model is typically used in this approach because of the large number of simulations that are required and because such models often explicitly include parameters such as the equilibrium climate sensitivity. The parameters that are varied from one climate simulation to the next typically include the equilibrium climate sensitivity α and other important determinants of the climate response to greenhouse gas forcing. The latter may include the effective vertical diffusivity of the ocean κ (which controls the rate at which heat anomalies penetrate into the deep ocean) or a parameter representing a range of possible aerosol forcings ε_{aer}.

Depending upon the study, the comparison between observations and model is performed either only in time (i.e., after integrating $T(x,t,\theta)$ and $T_{obs}(x,t)$ over the space coordinate x) or in both space and time. Also, depending upon the study, T and T_{obs} can represent either a scalar variable such as surface temperature, or a vector composed of several variables such as surface temperature, upper-air temperature and deep-ocean temperature.

A variety of statistics have been used to evaluate the agreement between model and observations for a given setting of the parameters θ. Knutti et al. (2002) assess the probability of a mean quantity being consistent with observations by calculating the probability of the observed change given the model simulation, its uncertainty and observational uncertainty. Forest et al. (2001; 2002) first calculate the residual mean square $\mathrm{r}^2(\theta, T_{obs}) = (\mathrm{T}(\theta) - \mathrm{T}_{\mathrm{obs}})^T C^{-1} (\mathrm{T}(\theta) - \mathrm{T}_{\mathrm{obs}})$ where C^{-1} is the inverse covariance matrix of internal climate variability estimated from control simulations with AOGCMs. As in optimal detection methods, this statistic measures residual variability after transforming the model response and observations so that the former is optimally detectable in the latter (Allen and Tett, 1999; see Appendix 9.A).The same square residual, but using a simple Euclidean measure, was used by Hegerl et al. (2006). The residual mean square is subsequently used to evaluate the likelihood of the given parameter choice. For example, assuming that noise is Gaussian, the likelihood can be evaluated by using the fact that the relative difference between any residual and the minimum residual $\Delta r^2/(r^2_{\min})$ will be distributed according to the F-distribution with m and ν degrees of freedom (Forest et al., 2002; 2006) where m is the number of free parameters in the model simulation, and ν is the number of degrees of freedom of the residual climate variability.

In either case, the end result is a *likelihood* function $p(T_{obs}\,|\,\theta)$ that describes how the likelihood of the observations changes as the parameters θ vary. This function, together with a prior distribution on the parameters, can be combined by means of Bayes theorem to obtain a posterior distribution $p(\theta\,|\,T_{obs})$ on the parameters. That is, one calculates $p(\theta\,|\,T_{obs}) \propto p(T_{obs}\,|\,\theta)\cdot p(\theta)$ where the product on the right is normalized so that the integral of $p(\theta\,|\,T_{obs})$ with respect to θ is equal to one. Finally, if one is interested in making inferences only about one of the parameters contained in θ, say the climate sensitivity, the posterior density function $p(\theta\,|\,T_{obs})$ is integrated over the ranges of other elements of θ to obtain the marginal posterior probability density function for the parameter of interest.

The prior distribution $p(\theta)$ that is used in this calculation is chosen to reflect prior knowledge and uncertainty (either subjective or objective) about plausible parameter values, and in fact, is often simply a wide uniform distribution. Such a prior indicates that little is known, a priori, about the parameters of interest except that they are bounded below and above. Even so, the choice of prior bounds can be somewhat subjective. In the case of climate sensitivity, several studies have used 0°C and 10°C as lower and upper bounds respectively (see Table 9.6.1).

An important consideration is the representation of the expected discrepancy between observed and simulated temperatures due to, for example, observational uncertainty or internal climate variability. This "noise" affects the width of the likelihood function, and thus the width of the posterior distribution. Noise properties must be estimated from either models or residual variability. Care is required because underestimates of noise will result in narrow posterior distributions that do not adequately portray the real uncertainty of parameters of interest.

Another important consideration is the choice of prior. The approach used in most studies is to place uniform priors on the parameters of interest, such as the equilibrium climate sensitivity. However, observable properties of the climate system do not necessarily scale with equilibrium climate sensitivity (Frame et al., 2005).Imposing a flat prior on an observable property, such as the climate feedback or transient climate response, is equivalent to imposing a highly skewed prior on the equilibrium climate sensitivity, and therefore results in narrower posterior likelihood ranges on the climate sensitivity that exclude very high sensitivities. Alternatively, expert opinion can also be used to construct priors, as done by Forest et al. (2002; 2006).Note, however, that expert opinion may be overconfident (Risbey and Kandlikar, 2002) and if this is the case, the posterior distribution may be too narrow. Also, the information used to derive the expert prior needs to be independent from the information that is used to estimate the posterior distribution. However, prior belief about the climate system tends to be shaped by observations of that system, and thus it is difficult to develop truly independent prior distributions.