Chapter 9: Understanding and Attributing Climate Change

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Date of Draft: 12 August 2005

Notes: This is the TSU compiled version

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Executive Summary

Evidence of the effect of external influences, both anthropogenic and natural, on the climate system has continued to accumulate since the TAR. Anthropogenic warming of the climate system is pervasive and can be detected in temperature observations taken at the surface, in the free atmospheric and in the oceans. It is very likely that the warming during the past half century is not solely due to known natural internal and external causes, and virtually certain that it is not due to natural internal climate variability alone. The warming occurred in both oceans and atmosphere, and took place at a time when non-anthropogenic forcing factors would likely have produced cooling and when the combined effect of known sources of anthropogenic forcing would have been very likely to produce a warming. It is likely that greenhouse gas forcing during the past half century caused greater warming than observed, with some of the greenhouse-induced warming being offset by cooling caused by aerosol and other forcings. 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 diverse parts of the climate system including the atmospheric circulation, the hydrological cycle and the cryosphere.

While many uncertainties remain, the physical consistency of the many different lines of evidence now available supports the conclusion that anthropogenic forcing is responsible for many of the changes in the climate system observed over the past century. Further evidence supporting this conclusion comes from many paleoclimatic studies indicating that the late 20th century was likely warmer than at any time during the past 1000 years, and that this warming took place at a time when non-anthropogenic factors would likely have produced cooling.

Our confidence in this assessment is increased by the preponderance of evidence that is now available from the large number of studies that have been performed. Many separate studies have now shown that climate change on global and continental scales during the past half century is consistent with anthropogenic forcing, and inconsistent with natural forcing or natural internal climate variability. These studies have used different detection and attribution techniques, applied them to different variables, used different climate models, and used different observational data sets. Each individual result has been assessed using a statistical method that limits the probability of obtaining spurious detection results by random chance to a low level. While these results are not completely independent, the chance that all results are spurious is very small.

Warming of the climate system is pervasive:
Anthropogenic warming of the climate system is pervasive and can be detected in temperature observations taken at the surface, in the free atmospheric and in the oceans

We now have 6 more years of high quality instrumental surface data. These observations show continued warming, and include extreme events such as the unprecedented 2003 European heat wave (Chapter 3). Each of the last six years numbers amongst the ten warmest years on record globally[1]; even though several of those years were La Niña years (which tend to lead to cooler global temperatures). The global mean temperature averaged over land and ocean surfaces during the period 1901–2004 increased by 0.75°C (Chapter 3). It is very unlikely that this change is due to natural causes (natural internal variability, or natural external forcing) alone.

Multi-signal detection and attribution analyses, which explore the joint roles of greenhouse gas and other anthropogenic and natural forcing agents, clearly indicate a response to greenhouse gas forcing. These results are robust to the choice of analysis technique and climate model, and have taken into account modelling, forcing, and observational uncertainty. The contribution of greenhouse gases to the global mean temperature change during the 20th century is likely to have been larger than the observed warming, with some of the greenhouse-induced warming being offset by cooling caused by aerosol and other forcings.

Further evidence has accumulated that there has been a significant anthropogenic influence on the temperature of the free atmosphere since radiosonde measurements became available in the late 1950s. The

[1] To be updated prior to publication. The six-year period referred to is 1999–2004. The 10 warmest years on record, in order, are 1998, 2002, 2003, 2004, 2001, 1997, 1995, 2000, 1999 and 1991.

observed pattern of tropospheric warming and stratospheric cooling can be attributed to the influence of anthropogenic forcing, particularly greenhouse gases and stratospheric ozone depletion, while the influence of greenhouse gas increases can be separated from other forcings. The combination of a warming troposphere and a cooling stratosphere has led to an increase in the height of the tropopause that is likely due to greenhouse gas and stratospheric ozone changes.

It is very likely that anthropogenic forcing has contributed to the warming of the upper several hundred meters of the ocean that has been observed during the latter half of the 20th century. Combined anthropogenic and natural external influences have also been detected in changes in the heat content of the upper 2000 meters of the ocean, although there is considerable uncertainty in observed heat content changes that arises from the sparse and infrequent sampling of many parts of the ocean.

Regional surface temperatures have likely been affected by anthropogenic forcing:
Several studies have shown that the anthropogenic signal in surface temperature changes is now detectable in continental and sub continental scale land areas. The ability of models to simulate many aspects of the temperature evolution on these scales and the detection of significant anthropogenic effects on individual continents provide compelling evidence for human influence on the climate. Although it is generally more difficult to attribute temperature changes to individual forcings in continental and sub continental regions, than to attribute global scale changes, human influence has been clearly identified in a variety of different studies in a number of disparate regions. As with the multiple lines of evidence available on global scales, 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.

Surface temperature extremes have likely been influenced by anthropogenic forcing:
Many indicators of impact-relevant surface temperatures, including the annual numbers of frost days, warm days and cold days, and other indicators such as the diurnal temperature range and the growing season length, show changes consistent with those anticipated from anthropogenic warming. Anthropogenic influence has been detected in some of these indices, including indices of extremely warm nights, cold days and cold nights. However, the same study was unable to identify anthropogenic influence in indices of extremely warm days on broad spatial scales. Nonetheless, there is evidence that anthropogenic forcing has substantially increased the risk of extremely warm summer conditions regionally, such as those that accompanied the unprecedented 2003 European heat wave.

Anthropogenic influence is becoming apparent in other parts of the climate system:
Recent decreases in Arctic sea ice extent are unlikely to be due to natural variability and agree with model-simulated anthropogenic changes. There is evidence of a decreasing trend in global snow cover, and widespread melting of glaciers, consistent with a widespread warming. Glaciers continue to retreat almost everywhere, with some exceptions such as in Northern Europe, where increases in precipitation appear to be increasing glacier volumes. These changes are generally consistent with those expected under anthropogenic forcing.

Significant trends in the Northern and Southern Annular Modes, which correspond to sea level pressure reductions over the poles, are likely to be partly related to human activity. While models reproduce the sign of the Northern Annular Mode trend, the simulated response is too small. Nonetheless, 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.

Evidence of the impact of external influences on the hydrological cycle is emerging. An anthropogenic influence on annual total precipitation has not been detected, consistent with 20th century climate simulations. However, there is some evidence that volcanic forcing has caused variability in global land-area total precipitation for physical reasons that are understood, although the model simulated changes are generally underestimated and not obtained with all models. Increases in heavy precipitation in some regions of the globe appear to be consistent with increases that are expected to occur with increasing anthropogenic forcing, although these changes are not yet clearly distinguishable from natural variability. A recent detection study attributed increased continental runoff observed in the latter decades of the 20th century to suppression of transpiration due to CO_2 -induced stomatal closure. Observed increases in the global frequency of drought in the second half of the 20th century have been reproduced with a model by taking anthropogenic and natural forcing into account. Observed changes in monsoon and Sahel rainfall appear to be related to changes in observed sea surface temperature conditions, which may in turn, have been affected by external forcing changes.

Climate models are able to reproduce climate conditions in the recent geologic past:
The same coupled climate models used to predict future climate have been used to understand and reproduce climatic conditions in periods of the recent geologic past, such as the Last Glacial Maximum and the Mid-Holocene. The Last Glacial Maximum features strong changes in external forcing relative to the present, while the Mid Holocene forcing exhibits a substantial change in seasonal forcing from today. Hence both provide a valuable test bed for climate models, particularly their feedbacks. 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.

Estimates of the climate sensitivity are now better constrained by observations:
There has been extensive study of climate sensitivity and feedbacks since the TAR, using a variety of models and data, including instrumental data, new reconstructions of the NH temperature record of the past millennium, and other periods in paleo-climate history such as the LGM. Estimates of sensitivity, their uncertainties and the ability of observable quantities to constrain these estimates are now much better understood. Climate feedbacks are very likely positive and the equilibrium climate sensitivity to CO_2 doubling very likely exceeds 1°C. Best estimates of climate sensitivity from various studies range from 1.5 to 4.5°C, but it remains difficult to constrain the upper limit. In contrast, the transient climate response is well constrained from observations and is very unlikely to be greater than 4.5°C per century in response to a 1% per year increase in CO_2.

Overall consistency of evidence:
Although uncertainties remain, the overall evolution of the climate over the 20th century is largely consistent with expectations based on the simulated response to external forcing. This consistency is apparent in many types of observations, including surface and free atmospheric temperature; top of atmosphere radiation anomalies; some large scale features of the atmospheric circulation; some aspects of precipitation and the hydrological cycle, ocean temperature and heat content; and sea-ice extent. Our confidence in our assessment of the role of humans in the recent climate evolution has increased considerably since the TAR because the anthropogenic signal has emerged in more aspects of the climate system, because models have been further improved, and because some apparent inconsistencies in the observational record (notably that for tropospheric temperatures) have been largely resolved.

Remaining uncertainties:
Estimates of radiative forcing change are still quite uncertain and limit the reliability of detection and attribution results. This is particularly the case for aerosol and other non-greenhouse gas anthropogenic forcings, and for low-frequency solar forcing. While uncertainties in the amplitude of the climate response to these forcings can be accounted for in detection approaches, uncertainties in the space-time pattern of the response can affect both detection and attribution results.

The effects of uncertainties in forcing and the simulated responses on detection and attribution results have not yet been fully explored. Furthermore, some potentially important forcings such as carbonaceous aerosols have not yet been considered in most formal detection and attribution studies, although their potential effects on the 20th century climate have been evaluated with a range of models. Consequently, it is possible that detection and attribution results are somewhat more uncertain than estimated. However, the robustness of surface temperature attribution results to forcing and response uncertainty has been explored with respect to a range of models, forcing representations and analysis procedures. Results suggest that it is very unlikely that these uncertainties are large enough to negate detection and attribution findings. There are greater uncertainties with respect to other variables, such as surface pressure and precipitation, because simulated responses to 20th century forcing change are in qualitative, but not good quantitative agreement with observations of these variables. The extent to which observational uncertainty, errors in forcing and errors in model physics contribute to these differences is not presently understood.

The estimates of internal variability used in detection and attribution studies, which are derived from climate models, have been increasingly scrutinized. Comparisons of model simulated variability with instrumental and proxy data have increased confidence that climate models simulate reasonable amounts of internal variability. However, uncertainties remain because the available observational records are influenced by external forcing, and because they are not long enough in the case of instrumental data, or precise enough in the case of proxy reconstructions, to provide accurate estimates of variability on decadal and longer time scales. Estimates of the internal climate variability of variables other than surface temperature, such as rainfall and ocean heat content, are more uncertain, in large part because of the limited observational resources available for these variables.

We have increased confidence that the temperature in the free troposphere is increasing, though uncertainty remains in both the radiosonde and satellite records, and this affects the confidence in the size of the estimated anthropogenic contribution to free atmospheric temperature change.

We are beginning to develop an understanding of changes in extremes, and an ability to assess forced changes in the frequency and intensity of extremes from observations and to assess changing risks of extremes. Incomplete global data sets for extremes analysis and model uncertainties still restrict detection studies of extremes.

We have improved our understanding of the causes of uncertainty in estimates of equilibrium climate sensitivity. These uncertainties, which include uncertainty in non-greenhouse gas forcing, particularly by aerosols, and ocean heat uptake as well as observational uncertainties in the preindustrial period, limit our ability to tightly constrain the upper confidence limit on likely values of the climate sensitivity. Furthermore, the upper limit for equilibrium climate sensitivity is much more difficult to constrain from observations of transient climate changes than the transient climate response.

9.1 Introduction

9.1.1 What do we mean by Climate Change and Climate Variability?

In order to understand climate change we must also understand natural climate variability. In the following we refer to climate variability that occurs in the absence of external forcing as *internal climate variability*. This variability originates from the chaotic, nonlinear interactions within and between the various components of the climate system. 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. These separate components involve processes that produce internal variability directly, but they also produce long timescale variability by integrating variability from the rapidly varying atmosphere (Hasselmann, 1976).

The internal climate variability is difficult to estimate because all climate observations are influenced, at least to some extent, by variations in external forcing. Estimates of internal variability can be obtained from models or observations if suitable assumptions are made. One such assumption is that the characteristics of the internal variability are not sensitive to small changes in external forcing. Another related assumption is that the statistics of the climate variability are constant, or stationary, when the climate is in equilibrium with a fixed forcing regime. Although it is not certain such assumptions are justified, there is little evidence to the contrary from either the climate history of the past few thousand years or from modelling studies.

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 defined in this way may be due to natural internal processes or external forcings, or to anthropogenic forcings. In this chapter, climate change refers mainly to change due to external forcings, while changes due to internal climate system processes are considered to be part of the climate's internal variability.

In order to understand the origins of observed climate variations, much research is performed to distinguish between natural internal climate variability and changes that are forced externally. Some external influences, such as changes in solar radiation and volcanism, occur naturally and can be considered part of the 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, and land-use change, are the result of human activity. Distinguishing between the effects of external influences and natural internal climate variability requires comparison between changes that are expected to result from external forcing and observed changes. The expectations are often quantified with climate models that are run with prescribed changes in external forcing. The most rigorous assessment which observed changes are due to forcing is obtained in detection and attribution studies.

9.1.2 What do we mean by Climate Change Detection and Attribution?

The concepts of climate change "detection" and "attribution" used in this chapter remain primarily as they were defined in the TAR (IPCC 2001; Mitchell et al., 2001). We use the same definition of attribution, which is discussed briefly below, but now use a more precise definition of detection than was given in the TAR. Specifically, we take *detection* to be the identification of one or more expected responses to changes in external forcing (for example, the climate response to greenhouse gas or volcanic forcing) in climate observations. This evaluation is often performed by estimating the amplitudes of the expected responses from *observations* and then determining whether these estimates are larger than might have been produced by random chance by internal climate variability alone. Note that the detection of an effect of an external forcing on the climate does not necessarily imply that the effect is important in the sense of having a major impact on the environment, biota, or human society. Consideration of such impacts is outside the scope of this chapter and Working Group.

Most 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. Climate modes are an appropriate tool for this purpose because they effectively integrate much our theoretical knowledge of the operation of the climate system. The patterns of expected response, which are commonly referred to as *fingerprints*, are usually derived from ensemble simulations with coupled climate models. Thus the hypotheses that are tested in most climate change detection studies have a firm basis in our understanding of the operation of the climate system.

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 result that is found to be significant, 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. The risk of such spurious detection can be diminished by appealing to corroborating lines of evidence that help to increase confidence in a result by producing a physically consistent view of the likely cause for the changes.

While the approach used in most detection studies assessed in this chapter is to determine whether observations exhibit the expected response to anomalous 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 risk of specific types of rare events.

Detection does not immediately imply attribution of the detected change to the assumed cause. As noted in the SAR (IPCC, 1996) and the TAR, 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

(a) detection as defined above,
(b) demonstration that the detected change is “consistent with the estimated responses to the given combination of anthropogenic and natural forcing”, and
(c) 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).

As discussed above, a key ingredient in all detection and attribution research is an estimate 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 anomalous 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 most often estimated from long control simulations produced with coupled climate models, which are used as a substitute for long records of observations of the unperturbed climate system. Therefore, the validation of model internal variability, through process studies and statistical comparisons with observed variability, provides important support for research that aims at understanding climate change (see Chapter 8). Detection studies often include specific steps to compare variability estimated from models with observed residual variability. Furthermore, although millennial forcing and proxy temperature reconstructions remain uncertain (see Chapter 6), comparison between climate simulations of the last millennium and proxy reconstructions also help to increase the confidence in model estimated internal variability.

A necessary part of any analysis that attributes climate change to specific causes is to determine whether the observed change is consistent with an estimate of the response to historical forcing, and to consider whether it can be explained by alternative hypotheses. The assessment of the consistency between an observed change and the estimated response to a hypothesized forcing is more challenging than the hypothesis testing that is performed in detection research (Appendix 9.A). In this case, the detection question is augmented with an evaluation of whether the amplitude of the hypothesized pattern of change estimated from observations is consistent with expectations. If so, the evidence for a causal connection is substantially increased. Note however, that consistency findings form only a part of the evidence that is used in attribution studies. Another key element is the evaluation of multiple lines of evidence, and the evaluation of the extent to which they are physically consistent.

A further part of an attribution analysis is ruling out that the observed change is not consistent with alternative explanations that exclude important elements of the given combination of forcings. This is done most reliably if the individual responses to the forcing agents that are hypothesized to have had an important influence on the climate are evaluated simultaneously and their influence on observations is separated from each other (see Appendix 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).

Model and forcing uncertainties are important considerations in attribution research. Ideally, the assessment of model uncertainty should include uncertainties in model parameters (as explored by multi-model ensembles, Chapter 8), and in the representation of physical processes in models. Such an assessment is not yet available. While research with that goal in mind is underway, computing constraints continue to limit progress in this area. Structural model uncertainties, such as those that result from the use of different types of parameterizations of atmospheric convection, are even more difficult to evaluate, although 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 forcing and that due to natural and anthropogenic aerosols (see Section 9.2), also remain difficult to evaluate. 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 climate during the latter half of the 20th century, are robust.

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).

Furthermore, for many variables, full detection and attribution studies are not feasible for a variety of reasons. Nonetheless, research that describes observed changes and offers physical explanations, for example, through numerical simulation with atmospheric, ocean or climate models, contributes substantially to our understanding of climate change and is therefore discussed in this chapter. This is particularly the case for variables such as rainfall which are less reliably modelled or observed, or respond less strongly to external forcing. For such variables, successful modelling of past changes contributes to our understanding of climate change and increases confidence in the ability of models to simulate future changes. Therefore, an important part of this chapter is to integrate the lines of evidence for detection and particularly for attribution of climate change.

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 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 accumulation of evidence since the TAR, and our growing understanding of uncertainties, are summarized in Table 9.1.1. Subsequent sections of this chapter will assess that evidence in detail. Here we briefly review the state of knowledge as it existed at the time of the TAR.

[INSERT TABLE 9.1.1 HERE]

The evidence that was available at the time of the TAR was broad and varied. Analysis of a range of several reconstructions of Northern Hemisphere surface temperature of the last several hundred to 1,000 years that had become available indicated that 20th century temperature changes were “unusual and unlikely to be entirely natural in origin”, even taking into account the large uncertainties in paleo-reconstructions. Using results from a range of detection studies of the instrumental record, and 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 took anthropogenic greenhouse gases and sulphate aerosols as well as natural forcing changes into account were found to be generally consistent with observations. In contrast, simulations of the response to natural forcings alone indicated that these may have contributed to the observed warming in the first half of the 20th century, but could not explain 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, to some extent, of temperature in the free atmosphere. Despite some problems, such as degeneracy between similar responses to different forcings, a distinct greenhouse gas signal was detectable whether or not other external influences were explicitly considered. 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 concentrations of greenhouse gases alone was found to be 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). These findings were reflected in the TAR’s strong conclusion on the anthropogenic contribution to climate change during the latter half of the 20th century.

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. However, there was also some concern about simulations showing a faster rate of warming in the mid- to upper troposphere over the satellite period than seemed consistent with the observations based on the satellite and radiosonde tropospheric temperature records that were available at the time.

A number of uncertainties remained at the time of the TAR. For example, the apparent discrepancies between the vertical profile of temperature change in the troposphere in the observations then available and models had been reduced, but not resolved. Also, large uncertainties remained in estimates of internal climate variability from models and observations. However, it was concluded that even substantially inflated (doubled or more) estimates of model simulated internal variance were 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. Large uncertainties were identified in anthropogenic aerosol forcing and response, and it was noted that some other anthropogenic factors, such as organic carbon, black carbon, biomass aerosols, and changes in land use, had not been included in detection and attribution studies. Uncertainties were also noted in reconstructions of solar and volcanic forcing, and in the magnitude of the climate response, particularly to solar forcing or in the model response to sulphate aerosol forcing. These uncertainties contributed to uncertainties in results from detection and attribution studies. Particularly, estimates of the contribution to the 20th century warming by natural forcings and anthropogenic forcings other than greenhouse gases showed some discrepancies with model simulations and were model dependent. These results made it difficult to attribute the observed climate change to one specific combination of anthropogenic and natural 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 discussing the scope of this chapter and by providing the reader with a road map of the following sections. As signalled by its title, the objective of this chapter is to understand observed climate changes that are reported in Chapters 3 to 5 as expressions of climate variability and externally forced climate change. Climate models, which integrate our theoretical understanding of the functioning of the climate system, are used to interpret those changes where possible. Our ability to interpret some changes, particularly for non-temperature variables, is limited at times by uncertainties in the available models, observations and external forcing estimates.

Thus, the scope of this chapter is wider than that of previous "detection and attribution" chapters in the SAR (Santer et al., 1996a) and the TAR (Mitchell et al., 2001). We do assess six more years of detection and attribution research, work that has ventured into some exciting and important new areas including detection of anthropogenic changes on regional scales, in extremes and in variables other than temperature. However, we also attempt to place the detection and attribution work in the context of a broader understanding of a changing climate by synthesizing information on climate forcing, variability and change from several other chapters.

The body of this chapter starts with a description of forcing changes and their uncertainties during the recent history of the climate system, and a brief summary of the climate response associated with these forcings (Section 9.2). The description of the response focuses on the similarities, differences and uncertainties in the climate's response to those forcing changes, 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 "top-down" methods. Such methods use the observed climate change to make inferences about the likely magnitude of uncertain external forcings to constrain estimates of uncertain forcings, for example, by identifying ranges of net aerosol forcing that yield simulations of 20th century climate change that are consistent with observed climate change given uncertainties in models and observations. This section thus attempts a synthesis that draws heavily on Chapters 2 and 8 of this report.

Section 9.3 follows with a synthesis of paleo-climate material that draws very heavily on Chapter 6. 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). These are periods for which we have relatively large amounts of proxy data, and a relatively good understanding of the climate's forcing regime. They are therefore periods that can be used to gain insights into the link between external forcing and climate. Moreover, the temperature history of the most recent one to two millennia is very important for placing modern secular climate change in context and understanding the climate response to natural forcing.

Section 9.4 focuses on understanding instrumental era temperature change. This section, which draws heavily on Chapters 3 and 8, reports recent developments in the detection and attribution of the effects of external influences on the surface and free atmospheric temperature, on global and regional scales, and using a range of techniques. This section extends and strengthens many of the lines of evidence that were described in the TAR and sharpens estimates of the human contribution to global surface temperature change observed during the past century. It also describes an emerging body of work that attempts to link external forcing changes with the possibility of changes in the intensity and/or frequency of extreme temperature events.

The objective of the following Section 9.5 is to assess large scale climate change in other variables and climate components. This section draws heavily on Chapters 3, 4, 5 and 8. Where possible, it attempts to identify links between related changes, such as those linking 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. We include in this section discussion of: (a) ocean climate changes, where there is evidence of an anthropogenic influence on ocean heat content; (b) atmospheric circulation changes, where there is evidence of an anthropogenic influence on the surface pressure distribution; (c) precipitation change, where, for example, there is evidence that natural forcing, may have influenced global mean precipitation rates during the 20th century, and that persistent sea-surface temperature variations may have influenced local and regional rainfall distributions, such as in the Sahel; (d) cryosphere change; and (e) sea-level change, both of which again show some evidence of an anthropogenic influence.

Finally, we conclude the chapter with a section that assesses studies that have attempted to use instrumental observations or paleoclimatic data to constrain estimates of key climate parameters such as its equilibrium sensitivity to the doubling of CO_2 concentrations. Many of these studies rely heavily on detection and attribution methods (in essence, they ask what range of parameters of interest produces simulations of the instrumental or millennium climate histories that are consistent with observations of these periods).

9.2 Radiative Forcing and Climate Response

Much research, including recent work that will be assessed in subsequent parts of this chapter, has used climate model simulations forced with estimates of historical forcing changes to understand and interpret climate change since the industrial revolution and during other parts of climate history, such as the last millennium. These results are affected both by uncertainties in model simulations and in observations. Uncertainties in simulations result from uncertainties in the radiative forcings that are used, and from model uncertainties that affect the simulated responses to the forcings. Both the uncertainty in external forcing and in the model response will affect the so-called “fingerprints” of climate change (see Appendix 9.A), and hence results of studies comparing simulated with observed changes.

Detection and attribution methods scale the response patterns to different forcings to obtain the best match to observations, thus errors in the magnitude of the forcing or the magnitude of a model’s response to anomalous forcing, should not affect detection results provided that the space-time pattern of the response is correct. However, for a model simulation to be considered consistent with the observations the scaling should be consistent with unity when bottom-up forcing estimates are prescribed in the model. 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.

In this section, we review the estimated ranges of forcing over the last 150 years, how models respond to different forcings, forcing uncertainties and implications for simulated responses and climate change detection and attribution.

9.2.1 Radiative Forcing Estimates Used to Simulate Climate Change

9.2.1.1 Summary of forcing estimates for the instrumental period

Radiative forcing is defined as “the change in net (down minus up) irradiance (solar plus long-wave; in W m^{-2}) at the tropopause after allowing for stratospheric temperatures to readjust to radiative equilibrium, but with surface and tropospheric temperatures and state held fixed at the unperturbed values” (see Chapter 2). Radiative forcing change since 1750 is reviewed in detail in Chapter 2. It describes estimated forcing change 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 H_2O increase from CH_4. These forcings and their estimated 90% confidence intervals (given in parentheses as a percentage of the central estimate) are well mixed greenhouse gases: 2.58 W m^{-2} (10%), stratospheric ozone: –0.10 W m^{-2} (40%), tropospheric ozone: 0.40 W m^{-2} (50%), total direct aerosol forcing: –0.2 W m^{-2} (100%) of which sulphate forcing (–0.4 W m^{-2}) and fossil fuel BC (+0.3 W m^{-2}) are largest, first indirect aerosol forcing from tropospheric aerosols: –1.2 W m^{-2} (30%), aviation induced contrails: 0.01 W m^{-2} (200%) and aviation induced ice clouds: (range from 0 to 0.05 W m^{-2}), surface albedo change: –0.03 W m^{-2} (235%), stratospheric H_2O: 0.13 W m^{-2} (200%), and low-frequency changes in solar radiation: +0.1 W m^{-2} (200%). While several of the IPCC AR4 models have included many of these (e.g., the long-lived greenhouse gases, direct and indirect sulphate aerosol, stratospheric and tropospheric ozone changes, black and organic carbon aerosols from fossil fuels, land use change effects on surface albedo, volcanic and solar irradiance changes) 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 bottom up studies is 1.53 W m^{-2} (67%) giving a likely range for the total net forcing of 0.48 to 2.58 W m^{-2} although a forcing close to zero is possible. The observed warming trend over the 20th century provides only a weak constraint on this large range of forcing estimates, as is further discussed in the following subsection. Nevertheless, it is the spatial-temporal pattern 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 "Top Down" estimates of aerosol forcing

"Bottom up approaches" to estimating forcing derive the forcing either from a model simulation based on estimated emissions or from a model simulation constrained by observations. These might yield a total net radiative forcing over the 20th century that could be close to zero or even negative (Boucher and Haywood, 2001). Net forcing close to zero allows high values of 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 unless the entire warming were due to natural internal variability, which is effectively ruled out if the natural variability in unforced climate models (or the amount of variability in paleo reconstructions that is not explained by external forcing) is taken as the measure of internal variability (Mitchell et al., 2001; see also Section 9.3.4 and 9.4.1.3).

Detection and attribution 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. Results from detection studies or related approaches can therefore be used to draw conclusions about the magnitude of uncertain forcings. This is done using "top down" methods (see Section 9.6 and Appendix 9.B). These 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; Andronova et al., 2005; Forest et al., 2001, 2002, 2005; Knutti et al., 2002, 2003; Harvey and Kaufmann, 2002; Stott et al., 2005b; see Table 9.2.1).

Different top-down approaches consider different external forcings as is summarized in Table 9.2.1. One type of top-down method is to determine the magnitude of the response to different external forcing agents from ranges of scaling factors for the climate change fingerprints that are derived from detection analyses (see 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 therefore errors in the model's transient sensitivity can be separated from errors in aerosol forcing in the model (assuming that errors in a model's sensitivity to greenhouse forcing and aerosol forcing are similar - see Gregory et al., 2002a; Allen et al., 2005; Table 9.2.1). By scaling up or down spatio-temporal patterns of response, 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 Earth System Models of Intermediate Complexity (see Table 8.8.2**)** or simple climate models to explore the range of forcings and climate parameters that yield results that are consistent with observations. Like detection methods, these approaches assess the fit of space-time patterns, or spatial means in time, to 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 essentially Bayesian approaches (Appendix 9.B), where prior assumptions for ranges of external forcing, which are often obtained from bottom up approaches, are updated using the observed climate change, yielding posterior distributions of external forcing magnitude that are consistent with observed climate change. As summarized in Anderson et al. (2003), the resulting range is generally narrower than the range from bottom up approaches due to the additional information used. However, the bottom-up approaches yield the forcing that results from a particular mechanism, whereas the top-down approaches yield the net forcing for the set of mechanisms that have a similar forcing pattern to that assumed a priori.

Studies attempt to separate greenhouse gas and aerosol effects by making use of either the hemispheric gradient in forcing, or the difference in the temporal variation in the relative 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. Global sulphur emissions (and thus, sulphate aerosol forcing) levelled off after 1980 (see Chapter 2.), 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).

A number of the ensemble-based studies cited in Table 9.2.1 find that the net aerosol forcing over the 20th century is negative but that to be consistent with the observed warming, it is estimated to be less than approximately –1.7 W/m^2. Harvey and Kaufmann (2002), who use an approach that focuses on the 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 top-down approaches (Table 9.2.1) suggest that total aerosol forcing may be somewhat weaker than the best estimates given in Section 9.2.1.1.

There is a range of uncertainties in these estimates. 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., Harvey and Kaufmann, 2002; Andronova and Schlesinger, 2001). However, the ratio of Northern to Southern Hemisphere forcing by industrial aerosols is not accurately known (see Section 9.2.2.2). Also, it is necessary to account for hemispheric asymmetry in tropospheric ozone forcing. Additionally, aerosols from biomass burning could be an important fraction of the total aerosol forcing although they have little hemispheric asymmetry. Furthermore, results will be only as good as the spatial or time pattern that is assumed in the analysis, and missing forcings may hamper the interpretation of results. Nevertheless, if interpreted carefully and considering the assumptions used, the top down results do provide valuable additional information on the possible range of forcings.

9.2.1.3 Radiative forcing of preindustrial climate change

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

Changes in the Earth's orbital parameters have been identified as the pacemaker of climate change on the glacial to interglacial timescale (e.g., Berger, 1988). These orbital variations, which can be calculated with strong confidence from astronomical laws (Berger, 1978), force climate variations by changing the seasonal and latitudinal distribution of solar insolation (see Chapter 6).

Solar insolation was similar to today during the LGM (21000 years ago). The major forcing for the Earth's climate at this time came from the presence of large ice-sheets in the northern hemisphere (Peltier 1994, 2004), and from the reduced atmospheric CO_2 concentration (185 ppm from recent ice core estimates, see Monnin et al 2001). These reduced CO_2 concentrations changed the radiative forcing by about –1.9 to –2.6 W/m^2 (best guess –2.2 W/m^2; see Chapter 6). Using several model simulations, the albedo forcing of the ice-sheet was estimated to range broadly from –2 to –4.3 W/m^2 (Chapter 6). These forcing changes are thought to be the result of ocean, vegetation and other feedbacks (Chapter 6). The radiative forcing from increased atmospheric aerosols (dust primarily) has been estimated to be about –1 ± 0.5 W/m^2 (Chapter 6).

Changes in orbital forcing during the mid-Holocene lead to a 5% increase of summer insolation in the northern hemisphere compared to the present. This corresponds to a negligible annual mean forcing of 0.011 W/m^2, and a 27 W/m^2 change in the magnitude of the seasonal cycle in forcing averaged over the northern hemisphere together with a -1W/m^2 change in the southern hemisphere.

Over the last millennium, changes in the Earth's orbit have had little impact on annual mean insolation. They lead to a reduction (increase) of summer (winter) insolation of 0.33 W/m^2 (0.83 W/m^2) at 45°N (Goosse et al. 2005), and a diminution of the magnitude of the insolation mean seasonal cycle of 0.4 W/m^2 in the northern hemisphere. Changes in insolation are thought to have arisen mainly from small changes in solar irradiance, although both timing and magnitude of past solar radiation fluctuations are highly uncertain (see Chapters 2 and 6; Gray et al., 2005; Lean et al., 2002). The other source of natural forcing 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 that have been used to force climate models either for the last two millennia (Crowley, 2000, updated), or the last 500 years (Robertson et al., 2001; see also Jones and Mann, 2004 for a review). 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, see Zorita et al., 2004 for a comparison; see also Chapter 6), and on the order of 35% 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 induce inter-decadal climate variability in reconstructions (see Section 9.3.4).

During the cool period of the Late Maunder Minimum (approximately 1675–1715), sunspots were generally missing, and solar irradiance is believed to have been smaller than before and after. This period will be used in Section 9.6 to discuss climate sensitivity; therefore we discuss its radiative forcing estimates here. 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 , Chapter 2), but with large uncertainties. This leads to a best estimate radiative forcing of –0.2 W/m^2 (–0.1 to –0.35 W/m^2 67% confidence interval; note that solar forcing from 1750 to the present is estimated having increased by 0.1 W/m^2, Chapter 2). Many radiative forcing changes, particularly those associated with industrialization, are very similar from the present to the Maunder Minimum as they are from the present to preindustrial (total forcing estimated of –1.53 W/m^2, see Section 9.2.1.2). This yields an approximate net radiative forcing of –1.63 W m^{-2} between the late Maunder Minimum and the present, with large uncertainties.

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

Our ability to distinguish between climate responses to different external forcing factors depends on the extent to which those responses are distinct. Figure 9.2.1 shows the zonal average temperature response to several different forcing agents over the last 100 years as computed in the PCM climate model (Santer et al., 2003a), while Figure 9.2.2 shows the zonal average temperature response to fossil fuel black carbon and organic matter, and to the combined effect of these forcings together with biomass burning aerosols computed with the CSIRO climate model (Penner et al., 2005). These figures illustrate that the modelled vertical and zonal signature of the temperature response varies with the forcing. Greenhouse gas forcing produces a warming in the troposphere, with cooling above, and somewhat more warming near the surface in the NH due to its larger land fraction (Figure 9.2.1d). 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 9.2.1c); thereby partially offsetting the greater NH greenhouse gas induced warming. The combined effect of tropospheric and stratospheric ozone forcing (Figure 9.2.1e) is to warm the troposphere and cool the stratosphere, particularly at high latitudes where stratospheric ozone loss has been greatest. These responses are distinct from those to the natural forcings. Solar forcing results in general warming of the body of the atmosphere (Figure 9.2.1a) with a pattern of surface warming that is similar to that of greenhouse gas warming, but differently from greenhouse warming, solar forced warming extends throughout the atmosphere. Volcanic forcing, on the other hand, results in warming in the lower stratosphere and near the surface at high latitudes, with cooling elsewhere. The net effect of all forcings combined is a pattern of Northern Hemisphere 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. Results obtained with the CSIRO model (Figure 9.2.2) suggest that black carbon, organic matter and biomass aerosols would further enhance the Northern Hemisphere warming that is shown in Figure 9.2.1e. On the other hand, indirect aerosol forcing from fossil fuel aerosols may be larger than the direct effects that are represented in the CSIRO model, in which case the Northern Hemisphere warming could be somewhat diminished.

[INSERT FIGURE 9.2.1 HERE]

[INSERT FIGURE 9.2.2 HERE]

One possible 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. Such forcing changes would be reflected in changes in surface radiation fluxes and surface warming. There has been continuing interest in this possibility (Stanhill and Cohen, 2001; Gilgen et al., 1998; Liepert, 2002). Called "global dimming", this phenomena has recently reversed since about 1990, with surface solar radiation fluxes as inferred from a combination of ISCCP data and model-results now apparently increasing and top-of-atmosphere ISCCP fluxes decreasing (Pinker et al., 2005; Wild et al., 2005; Wielicki et al., 2005). The consistency between the IPCC AR4 models and these more recent analyses is shown in Figure 9.2.3 where the top-of-atmosphere (TOA) outgoing flux from the models is compared with that measured by the ERBS satellite (Wong et al., 2005) and inferred by ISCCP. Overall, the model range and the observational range encompass each other closely, and both have small downward trends, which appear slightly smaller in most models than in the observations. Differences between simulated and satellite trends are significant for only a few models.

[INSERT FIGURE 9.2.3 HERE]

The TOA flux changes displayed in Figure 9.2.3 are sensitive to changes in cloudiness (with decreasing cloudiness consistent with decreasing TOA fluxes) and changes in scattering aerosols. TOA fluxes are not as sensitive to changes in absorbing aerosols although the surface flux anomalies would be sensitive to both changes in absorbing and scattering aerosols as well as changes in cloudiness. Changes to the surface irradiance may be examined using the ISCCP FD data (Zhang et al., 2004). Romanou et al. (2005) compare the regional surface irradiance changes in a multi-model mean from a subset of the AR4 models shown here. They demonstrate that some of the regional trends present in the data are indeed present in the models, though the regional trends in the model-mean data set are smaller than those in the satellite-derived data set.

Figures 9.2.1 and 9.2.2 also show that the spatial signature of a climate model's response is seldom like that of the forcing. This comes about because climate system feedbacks vary spatially. 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., Rotstayn and Penner, 2001; Mitchell et al., 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 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). Therefore, the strength of the feedbacks within a climate model, and indeed, whether those that are important to determining the response to a given forcing are included, will help determine the spatial signature of its response. 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 as opposed to well mixed greenhouse gases, tend to have a more distinct spatial pattern of forcing, and can therefore also be expected to have some locally distinct response features.

The pattern of response to a radiative forcing can also be altered quite substantially if the atmospheric circulation is affected by the forcing. Modelling studies and data comparisons suggest that volcanic aerosols (e.g., Kirchner et al., 1999; Shindell et al., 2001b, Stenchikov et al., 2005), 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., 2001b) 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., 2005); leading to Eurasian winter warming that may reduce the overall cooling effect of volcanic eruptions on annual averages, particularly over Eurasia.

The temporal evolution of the different forcings (see Chapter 2, Figure 2.9.3 generally helps to distinguish the responses to the given forcings. For example, Santer et al. (1996b, c) 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 temporal 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.

9.2.2.2 *Uncertainty in the spatial pattern of response.*

Most detection methods identify the magnitude of the space-time pattern of response that is optimally consistent with the observations. The former are obtained by forcing a climate model with a given pattern and amplitude of 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; Rotstayn and Liu, 2003; Menon et al., 2002; Chuang et al., 2002). 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; Menon et al., 2002; Lohmann and Lesins, 2002). 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 that any can be excluded, 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. For example, there are uncertainties 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 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. 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) is significantly different from that 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 that 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.

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, but are not necessarily supported by current understanding (Lean et al., 2002; Foukal et al., 2004, Gray et al., 2005). 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; Amman et al., 2003).

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. Any given model simulation will normally use forcings that have been developed from bottom-up estimates. The best-quantified forcing is perhaps that from the most recent large volcanic eruption (i.e., Pinatubo in 1991), though opinions differ with respect to it's usefulness in quantifying responses (Wigley et al., 2005a; Frame et al., 2005).

It has been shown that for most forcings, the global average surface temperature response per unit forcing is similar within a given climate model (to within approximately 40%) (Hansen et al., 1997; Forster et al., 2000; Rotstayn and Penner, 2001; Gregory et al., 2004b). To the extent that the patterns are similar, it is possible to use methods that scale the response patterns to different forcings in detection work to obtain the best match to observations. Thus errors in the magnitude of the forcing or the magnitude of its response to anomalous forcing (which is approximately, although not exactly, a function of sensitivity), should not affect detection results provided that the space-time pattern of the response is correct. 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. Attribution studies, on the other hand, 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 scaling should remain consistent with the uncertainty bounds from bottom-up forcing estimates.

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 somewhat 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.

Increasingly, different models are used to include some of the uncertainty in the pattern of response (e.g., Hegerl et al., 2000; Gillett et al., 2002c; Santer et al., 2005b, Allen et al., 2005; Stott et al., 2005b, Zhang et al., 2005). This approach accounts for both forcing uncertainty, since usually, different model simulations apply different combinations of forcings, and model uncertainty. However, the limited sample of forcings and models used allow only for a limited exploration of these uncertainties. Some recent modelling studies have explored uncertainties associated with some model parameters (Stainforth et al 2005; Murphy et al 2004).

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. For example, solar and volcanic forcings correlate over the 20th century (for example, North and Stevens, 1998).

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 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 estimates of the response of a model to greenhouse gas forcing are not very sensitive to differences in simulated responses to aerosol forcing (Hegerl and Allen, 2002). In order to understand whether the range of response patterns used in detection studies adequately captures the range expected from studies of climate forcing, Stott et al. (2005b) consider the robustness of attribution results to uncertainties in patterns of aerosol forcing. They conclude that a key constraint is the temporal evolution of the global mean and hemispheric temperature contrast of the pattern of response to aerosol forcing (see Sections 9.2.4, 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, the uncertainties are large for many non-greenhouse gas forcings. Bottom up estimates of 20th century radiative forcing change suggest that it is very likely that the total anomalous forcing is positive but less than 2.58 W/m^2. Top-down studies, which use methods closely related to those used in climate change detection research, indicate that the magnitude of the net aerosol forcing is very likely less than –1.7 W/m^2.

In order of importance based on magnitude of forcing and inherent uncertainty, it is most important to better constrain those forcing agents that are associated with the indirect effects of aerosols, followed by those associated with the direct effects of biomass aerosols and fossil-fuel related aerosols. The large uncertainty in total aerosol forcing and response presently inhibits attempts to accurately infer the climate sensitivity from observations (Section 9.6). It also increases uncertainties in results that attribute cause to observed climate change (Section 9.4.1.4), and is in part responsible for differences in projections of future climate change (Chapter 10). Forcings from black carbon, fossil fuel organic matter, and biomass burning aerosols, which have not been considered in most detection studies performed to date, are likely small but with large relative uncertainties. Nonetheless, the distinct spatial pattern 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. For example, changes in the solar forcing over time need better quantification. While the 11-year solar forcing cycle is well-documented, lower-frequency forcing variations have been recognized as highly uncertain. In contrast, the timing and duration of forcing due to aerosols that are 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. For example, 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). Note that a spurious correlation between the climate responses to solar and volcanic forcing over parts of the 20th century (North and Stevens, 1998; North et al., 1998) can lead to misidentification of one versus the other as in the recent study by Douglass and Clader (2002). Therefore, results based on longer periods provide more reliable attribution of climate variability associated with natural forcing.

We conclude that due to differences in space-time patterns of greenhouse gas forcing from other anthropogenic forcing, a separation of the latter from other anthropogenic forcing, such as sulphate aerosol forcing (especially if the latter is taken to include the indirect forcing associated with fossil-fuel related aerosols), should be relatively robust. In contrast, as discussed above, the climate response and time evolution of other anthropogenic forcings is more uncertain, making simulated climate response and detected signals more difficult to interpret.

9.3 Understanding Pre-Industrial Climate Change

9.3.1 Why do we Consider Pre-Industrial Climate Change?

Understanding past climate variations is important for understanding both present and future climate change. The Earth system has undergone large-scale climate changes in the past (see Chapter 6). These changes resulted either from internal feedbacks in the climate system, or from the response of the climate system to natural external forcings. Our motivation for studying pre-industrial climate change is to better understand the variability and the major feedbacks of the climate system in the absence of anthropogenic forcing, and the sensitivity of the climate to external forcing. Past periods offer the potential to increase the information available from the instrumental record, which is too short to fully understand climate variability on interdecadal and secular timescales, and is affected by strong anthropogenic as well as natural external forcings.

Here we restrict the discussion mainly to the observations and simulations of the last millennium, because this period is vital to place the last 140 years in a broader context. 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).

9.3.2 Records of Past Climate Changes

Prior to the instrumental era, it is necessary to use indirect indicators ("proxy data") to infer past climate variations (see Chapter 6). These indicators include tree ring width and density, fossil corals, oxygen isotope ratios in ice core or marine cores pollen counts in sediment layers, and sulphate deposition on ice caps. These data provide information on past changes in climate, such as in temperature, precipitation, or other indicators that reflect changes in large-scale patterns of atmospheric and oceanic circulation which may have contributed to past climate changes. We give a very brief description of the paleo records here. A corresponding description of the relevant forcing situations can be found in Section 9.2.1.3. More complete information about these data and their uncertainties, and on the evolution of past climate changes, is provided in Chapter 6.

Compilations of proxy temperature records show that the Earth has warmed rapidly over the past 1000 years or so (see Chapter 6 and the reviews by Jones et al., 2001 and Jones and Mann, 2004). The Northern Hemisphere was likely cooler than the late 20th century during the first half of the millennium and cooled a further 0.5°C to –1°C below the 20th century mean during the period 1500 to 1900. This long term tendency is punctuated by several climatic events of about 0.25°C when averaged over the Northern Hemisphere (Figure 6.8). New reconstructions suggest larger variations than contained in the Mann et al. (1999) reconstruction (see Chapter 6), but large uncertainties remain about the inter-decadal and inter-century variability recorded by these reconstructions. Uncertainties in the amplitude of the reconstructed temperature variability arise from different sources depending on the proxy data and methods that are used. New methods for processing proxy data, such as from tree rings, better preserve low-frequency variability, but some uncertainty remains as to how to best calibrate proxy data to instrumental temperatures (e.g., von Storch et al., 2004; Mann et al., 2005; see also Chapter 6), although recently developed techniques that recover low frequency variability produce reconstructions that appear to be in good agreement (Moberg et al., 2005; Hegerl et al., 2005). 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 6.8).

In addition, high resolution records tell us how the interannual variability has evolved. For example, Cobb et al (2003), using fossil corals, has attempted to extend the record of ENSO back through the last millennium. They find that ENSO 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 12^{th} and 14^{th} centuries when there may have been significantly less ENSO variability, a period during which there were also cooler conditions in the North East Pacific (Mac Donald and Case, 2005) and evidence of droughts in central North America (Cook et al., 2004). 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).

The magnitude of natural climate fluctuations is larger further back in time (Chapter 6). Relatively high quality global reconstructions exist for the mid-Holocene and last glacial maximum (LGM) as part of the BIOME6000 project (Prentice and Webb, 1998; Prentice and Jolly, 2000). The climate was much colder and drier during the LGM with a strongly damped hydrological cycle relative to our modern climate as is indicated by the extensive tundra and steppe vegetation that existed during this period. Most LGM proxy data also suggest that the tropical oceans were colder by about 2°C than at present, and that the frontal zone 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 (De Vernal et al., 2005). 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 found in the now dry sub-Saharan regions indicate that wetter conditions prevailed there that resulted from enhanced summer monsoons (see Braconnot et al 2004 for a review). In addition, there is now also a wide range of proxies that extend the record of ENSO back in time for this period (Chapter 6). These data suggest that ENSO variability appears to have been weaker than it is today prior to approximately 5,000 years before present (Moy et al., 2002 and references therein; Tudhope and Collins, 2003).

9.3.3 What can we Learn from the LGM and the Mid-Holocene?

Several simulations with models of intermediate complexity (EMICS), as well as more comprehensive general circulation models, are available for the LGM and the mid-Holocene as part of PMIP (Joussaume et al., 1995). This includes 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.

Several new AOGCM simulations of the LGM have been produced since the TAR. These simulations generally simulate a global cooling of –3.1°C to –5.2°C in response to greenhouse gas and ice sheet forcing (Masson-Delmotte et al., 2005), 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). The range of simulated temperature change simulated by the coupled models reflects differences in their sensitivity to anomalous greenhouse gas forcing and changes in the specification of ice-covered regions (Taylor et al., 2000). Changes in greenhouse gas concentrations may account for about half of the 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. Some simulations are in qualitative agreement with the reduction of deepwater ventilation suggested from marine proxy data (Duplessy et al., 1988) and by a new reconstruction of the north Atlantic ocean (Kageyama et al., 2005). Inclusion of changes in vegetation appears to improve the realism of LGM simulations. Colder temperatures are simulated in Eurasia where forests retreat, allowing snow to have a larger impact in reflecting solar radiation back to space (Crowley and Baum, 1997; Levis et al., 1999; Wyputta and McAvaney, 2001). Inclusion of the physiological effect of the CO_2 concentration on vegetation has a non-negligible impact (Levis et al., 1999) and needs to be included to properly represent changes in global forest (Harrison and Prentice, 2003).

Coupled simulations of the mid-Holocene produce an amplification of the mean seasonal cycle of temperature of 0.55°C to 0.71°C. This range is slightly smaller than that of PMIP1 (0.55°C to 1.16°C) with atmosphere only models due to the thermal response of the ocean (Braconnot et al., 2000). Changes in the ocean circulation have strong seasonal features with an amplification of the SST seasonal cycle of 1 to 2°C in most places within the tropics. This amplification alters land-sea contrasts and influences regional climate change (Zhao et al., 2005). In particular, the late summer and autumn warming of the surface ocean in response to the insolation forcing enhances the land-sea contrast and affects both the Indian and African monsoons, favouring a late monsoon retreat. In Africa, changes in annual mean precipitation simulated by coupled models over West Africa 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 for future model projections (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 (Braconnot et al., 1999; Ganopolski et al., 1998). This has been confirmed in recent results (Levis et al., 2004), together with the role of the soil feedbacks. Changes in the land surface cover, and thus also surface albedo and surface roughness, introduce an annual mean climate forcing. The signal from this feedback is superimposed on the seasonally varying mid-Holocene insolation forcing. 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. A comparison with vegetation reconstructions reveals that coupled models reproduce most observed features, but have a tendency to overestimate the mid-continental drying in Eurasia (Wohlfart et al., 2005). This point needs to be considered when analysing future climate changes in these regions from model scenarios.

Several studies have now attempted to analyse changes in interannual variability for paleo periods relative to the present. 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). 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; Kitoh and Mukarami, 2002; Liu et al., 2000; Otto-Bliesner et al., 2003). A key element of the ENSO response is the Bjerknes feedback mechanism. Bjerknes (1969) argued, on the basis of observations, that if the East Pacific warms slightly, the trade winds weaken causing a deepening of the thermocline and consequently further warming. The increased mid-Holocene solar heating in boreal summer leads to more warming in the western than eastern Pacific, which sets off the Bjerknes feedback such that there is a cooling in the east. 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, even though 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 (Glastone et al., 2005).

9.3.4 What can we Learn from the Past 1000 Years?

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. These external forcing changes are, in general, small relative to those from the mid-Holocene and LGM to the present. Nonetheless, the climatic responses to forcing, together with internal variability, produce several well-defined climatic events, such as the late Maunder Minimum. 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, variability in reconstructions that is not explained by external forcing can provide an estimate of the magnitude of internal climate variability on interdecadal and secular timescales.

9.3.4.1 Evidence of external influence on the climate over the past 1000 years

Several transient simulations covering the last millennium have been performed using a range of models, including some runs with CGCMS (e.g., Crowley, 2000; Bertrand et al., 2002; Gerber et al., 2003; Bauer et al., 2003; Zorita et al., 2004, see also Gonzalez-Ruoco et al., 2003; Jones and Mann, 2004; Goosse et al., 2004, Tett et al., 2005). All of these simulations use reconstructions of historic solar and volcanic forcing variations (Figure 6.10). Despite the differences between the various experiments, which result from the use of different models and uncertainty in the various natural forcing reconstructions (see Section 9.2), they display some common characteristics in the evolution of the Northern Hemisphere annual mean temperature, consistent with the broad features shown by data. There is a general cooling from the beginning of the millennium until the 19th century that ranges from approximately 0.5°C to 1°C, depending upon the simulation considered, which is followed by a rapid warming. All simulations suggest that the time of the late Maunder Minimum (around 1675 to 1715) was one of the coldest periods of the millennium, which is qualitatively similar between models and proxy reconstructions, and that the 20th century was warmer than the last 800 years. The rapid warming at the end of the millennium leads also, with the possible exception of Zorita et al. (2004), to an absolute maximum during the last decades of the 20th century even compared to the warm 11th and 12th century. Beyond these broad agreements, both simulations of last millennium and proxy reconstructions show differences, particularly in the magnitude of past climate variations. The magnitude of the simulated climate variations varies between models due to differences in forcing and model formulation, and in reconstructions due to different proxies included and reconstruction techniques (Section 6.5.4). Nevertheless, all of these simulations 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, 2005a; Tett et al., 2005), stressing the impact of human activity on the recent warming. A new detection and attribution analysis of the reconstructions indicates that 20–50% of the warming of the first half of the 20th century can now be attributed to greenhouse gas emissions (Hegerl et al., 2005a).

Given the wide range of uncertainties on the magnitude of the Northern Hemispheric temperature trend from proxy data, it is not yet possible to use the proxy data to fully assess the realism of these simulations. Part of the variation in the magnitude of the simulated response can be attributed to differences in climate sensitivity, which varies by a factor of two between different climate models (Chapter 8). 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 reasonable. Differences in the simulation of internal variability between models could have contributed to the apparent differences between their simulations of Northern Hemisphere mean temperature, but 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; Goosse et al., 2004; Hegerl et al., 2003, 2005a). Other sources of uncertainty include the model 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, 2005).

9.3.4.2 Role of volcanism and solar irradiance

The role of volcanic forcing has been identified by comparisons between proxy records and model simulations. Volcanism manifests itself by large and rapid decreases of annual mean temperatures during the years following a major eruption. In addition, changes in the frequency of large eruptions result in climate variability on decadal to centennial time-scale (Crowley, 2000; Briffa et al., 2001; Bertrand et al., 2002; Bauer et al., 2003; Table 9.3.1). Hegerl et al. (2003, 2005a) detect a highly significant response to volcanic and greenhouse gas forcing in a number of proxy reconstructions, using a multi-regression approach (see Section 9.4, Appendix 9.A.1) to simultaneously estimate the response to anthropogenic and natural climate forcings from several paleo-reconstructions of Northern Hemispherically averaged temperature (Figure 9.3.1; Table 9.3.1). Also, modelling studies suggest volcanic activity has a dominant role in explaining the Maunder Minimum cold conditions (Yoshimori et al., 2005; Andronova et al., 2005). However, Rind et al. (2004) and Shindell et al. (2003) estimate from model simulations that the dominant forcing for the Maunder Minimum cold climate state were changes in solar radiation. These uncertainties in interpretation also reflect 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.

Tropical variability may also be affected by volcanism. 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). This study also suggests that, as in earlier periods (Section 9.3.3), the response of the mean state of the tropical Pacific depends upon differences in surface heating in the eastern and western Pacific. Other mechanisms could have a role, as has been shown for ENSO variations reported for the modern climate (see Chapters 2 and 8). However, substantial uncertainties remain as to whether and how ENSO changes in response to volcanism.

[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, 2005a), although the effect of solar forcing is detected over parts of the 20th century in some time-space analyses (see Section 9.4.1.4). We conclude that the preindustrial temperatures estimated by proxy data very likely show effects from natural external forcing, such as changes in solar radiation and explosive volcanism.

There is some evidence that the response to external forcing may influence modes of climate variability. 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, 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 Section 8.7.1.3; 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), as do reconstructions of this pattern for the northern hemisphere (Luterbacher 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 late Maunder Minimum (Shindell et al., 2001a; Section 9.2.2).

9.3.4.3 Other forcings and sources of uncertainties

In addition to forcing uncertainties discussed above, there are a number of other uncertainties that affect our 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 (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 (Chapter 2; Section 9.2.1.1). Model simulations s (e.g., Betts, 2001; 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 Mann, 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 thermohaline circulation in response to a decrease in solar irradiance (Cubasch et al., 1997; Weber et al., 2004; Goosse and Renssen, 2004). Because of the large oceanic inertia and the long time scales involved in the renewal of deep-water masses, the climate state at a particular period could be influenced by the previous climate evolution. A delayed response to 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). If this mechanism is valid, it would imply that the Southern Ocean has only partly responded to present-day changes in radiative forcing and large temperature increases could therefore be expected in the future (Weaver et al., 2000; Goosse and Renssen,. 2001).

9.3.5 Summary

The discussion above illustrates that climate models are important tools for interpreting past climate changes that are indicated by paleo records. Considerable progress has been made since the TAR 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, since it is now possible to use robust features from proxy data to assess paleo-climate simulations (see Chapter 6). Progress has been made in our understanding both of changes in the mean climate state and of changes in the short-term variability.

We have also substantially increased our understanding of the past thousand years, although uncertainties remain in the Northern Hemisphere temperature and natural forcing reconstructions. The major climatic events of the last millennium are quite well established at the scale of the northern hemisphere, but 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, and possibly for the last 2000 years (see Chapter 6). It has also been shown that the response to natural external forcing has significantly influenced the evolution of Northern Hemispheric temperature, with pronounced volcanic and possible solar effects, which contribute substantial interdecadal variability. 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 five 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 during the period 1901–2004 increased by 0.75°C (Chapter 3; approximately 0.6°C per century if approximated by a linear trend) and 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 (Chapter 3). A larger number of proxy reconstructions from paleo data than were available for the TAR indicate that Northern hemisphere mean temperatures of the late 20th and early 21st centuries are very unusual and likely unprecedented in the context of the last 2000 years (Chapter 6). Global mean warming since 1900 has not been a gradual smooth increase as would be expected if global mean temperatures correlated simply with 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 as rapid as that seen since the mid 1970s 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 (Figure 3.2.6). Since the mid-1970s, land regions have warmed at a faster rate than oceans in both hemispheres (Figure 3.2.8) and warming over the southern hemisphere was smaller than that over the northern hemisphere during this period (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 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 include increasing greenhouse gases, the effects of aerosols and 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.

[INSERT FIGURE 9.4.1 HERE]

Figure 9.4.2 displays the evolution of the spatial distribution of bi-decadal mean observed near surface temperature over the 20th century relative to the 1945–1974 mean (left hand column) and its simulation by an ensemble of 10 climate models that include anthropogenic and natural forcing (right hand column). The similarity between the general evolution of the warming in observations and models that include anthropogenic and natural forcing is compelling. 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 9.4.1c suggests that current models generally simulate large scale natural internal variability quite well. Section 9.4.1.3 will assess the variability of near surface temperature observations and simulations.

[INSERT FIGURE 9.4.2 HERE]

Global mean and hemispheric scale temperatures are controlled by external forcings on multi-decadal time scales. Stott et al. (2000) analysed an ensemble of integrations of HadCM3 (see 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.

The global mean warming observed since 1970 is captured by many different climate models when they are forced with external combinations of forcing that include anthropogenic forcings. These include simulations by the HadCM3 model including time varying changes in well-mixed greenhouse gases, the direct and first indirect effects of sulphate aerosols, and changes in tropospheric and stratospheric ozone (Tett et al., 2002), simulations by the GFDL R30 model including well-mixed greenhouse gases and the direct effect of sulphate aerosols (Broccoli et al., 2003), simulations by the GFDL CM2.0 and CM2.1 models including well-mixed greenhouse gases, tropospheric and stratospheric ozone, black and organic carbon, land cover changes, volcanic aerosols and solar irradiance changes (Knutson et al., 2005), and simulations by the NCAR PCM model including well-mixed greenhouse gases, the direct effect of sulphate aerosols and changes in tropospheric and stratospheric ozone (Meehl et al., 2004). In all these models, a good simulation of warming since 1970 is obtained when just anthropogenic forcings are included in the model and 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 a greater degree of uncertainty over the causes of early 20th century warming than the recent warming. Whereas some simulations imply a greater role for solar forcing than other forcings before 1950 (Meehl et al., 2004), other studies imply that volcanic forcing (Broccoli et al., 2003) or natural internal variability (Delworth and Knutson, 2000) could be more important. There is also likely to be an early expression of greenhouse warming in the early 20th century (Tett et al., 2002; Tett et al., 2005; Hegerl et al., 2003). Nozawa et al. (2005) detect a significant natural contribution to early 20th century warming and find that natural forcings are probably more important than anthropogenic forcings during this period. 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. Andronova and Schlesinger (2000) proposed that a natural oscillation over the North Atlantic and its adjacent land areas with a period of 65–70 years (Schlesinger and Ramankutty, 1994) could account for some of the evolution of global mean temperatures during the instrumental period, 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. In contrast, Nagashima et al. (2005) find that carbonaceous aerosols are required for the MIROC model to provide a statistically consistent representation of observed changes in near-surface temperature in the middle part of the 20th century.

The ability of climate models 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 persuasive evidence for the influence of humans on global climate. Nonetheless, given the large uncertainties in aerosol forcings, agreement could have been obtained fortuitously as a result of, for example, balancing too much (or too little) greenhouse gas warming by too much (or too little) aerosol cooling, and there is some evidence for a possible negative correlation between models' sensitivity and their total forcing over the century. 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 likely responses to individual forcings. 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 MSU temperatures (Santer et al., 2003a). Additivity is less likely to hold for regional responses. Meehl et al. (2003) found that 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.

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 9.4.1c. 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 (Stone et al., 2005a). 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 will not be entirely removed by 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 most models display similar variance to the observed variance on the decadal to inter-decadal time-scales important for detection and attribution.

[INSERT FIGURE 9.4.3 HERE]

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. However, a common experimental design for coupled model simulations of past climate change has been to make ensembles of transient simulations only until 2000, thereby limiting the number of analyses which consider 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 (Gillett et al., 2002c; Tett et al., 2002; Zwiers and Zhang, 2003; IDAG, 2005; Stott et al., 2005a; Stott et al., 2005b; Zhang et al., 2005; Allen et al., 2005). An analysis of global mean data from the IPCC AR4 20C3M models by Stone et al. (2005a) and an analysis using an EBM to infer the likely observed global mean responses to natural and anthropogenic forcings (Stone et al., 2005c) 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. In another study, Kaufmann and Stern (2002) detected a significant human influence on near-surface temperatures by analysing the lagged covariance structure of hemispheric mean temperatures.

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. They analysed a variety of simulations in which the model was forced with changes in natural forcings from changes in solar irradiance and stratospheric aerosol due to explosive volcanic eruptions, changes in well-mixed greenhouse gases and changes in other anthropogenic forcings including tropospheric and stratospheric ozone changes and the direct and first indirect effects of sulphate aerosol. By analysing observed and modelled near-surface temperature changes using an optimal detection methodology, they found that they could detect 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 chaotic variability on signal estimates from finite-member ensembles, could lead 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 low bias is likely to be greater for weak forcings such as changes in solar output (Allen and Stott, 2003; Stott et al., 2003a). Stott et al. (2003b) showed that the climate response to solar forcing could be significantly underestimated by HadCM3, a result that was also found by Crooks (2004) for temperature changes in the free troposphere. However, Gregory et al. (2004b) showed that HadCM3 has a lower sensitivity to solar forcing than greenhouse forcing. In contrast to the results from the 20th century, analysis of pre-industrial temperatures (Hegerl et al., 2003) suggests an overestimate of the response to solar forcings in models and suggests only relatively a small contribution of solar forcing to the early 20th century warming. Therefore, the evidence for the role of solar and volcanic forcing at present is mixed. Spurious correlations between these forcings during the 20th century (North and Stevens, 1998) and the uncertain time-evolution of solar forcing (see Section 9.2.1.3) have so far hindered a reliable estimate of the observed response to solar forcing. However, even with large amplifications of the HadCM3's response to solar forcing, the conclusion of the TAR still holds that most of the warming over the last 50 years of the 20th century is attributable to increasing greenhouse gas concentrations. On the whole, taking account of sampling uncertainty (as developed by Allen and Stott, 2003, and implemented by Stott et al., 2005a 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 and noisy signals.

Stott et al. (2005a) compared HadCM3 results over the 20th century with those obtained using the PCM and GFDL R30 models. They found consistent estimates for the greenhouse gas attributable warming of 0.7 to 1.3°C offset by cooling from other anthropogenic factors (associated mainly with cooling from aerosols) of 0.2 to 1°C and a small net contribution from natural factors over the century of between –0.1 and 0.1°C (first three sets of bars, Figure 9.4.4 middle panel). Scaling factors for the three forcings are shown in Figure 9.4.4 (top panel). A similar analysis for the MIROC model finds a slightly larger warming contribution from greenhouse gases of between 0.8 and 1.6°C offset by a cooling of –0.4 to 1.1°C from other anthropogenic factors and a very small net natural contribution (fourth set of bars in Figure 9.4.4). In all cases, the 5th percentile of the warming attributable to greenhouse gases is greater than the observed warming over the 20th century as a whole and over the last 50 years of the 20th century (Figure 9.4.4, middle and bottom panels). These results, obtained from four different climate models, with different sensitivities and forcings, are 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 9.4.4 HERE]

Not all models have the full range of simulations required. In these cases the likely pattern of a model's response to each forcing can be diagnosed, as was done by Crooks et al. (2005) by regressing the all-forcings model response onto the individual forcing time series to produce a spatial pattern of response for each forcing. The patterns are modulated in time by the magnitude of the forcing to produce time dependent response signals. The magnitude of these diagnosed signals can then be inferred from observations using optimal detection. Crooks et al. (2005) estimate attributable warming rates for HadCM3 that are consistent with those obtained using the full set of simulations (compare fifth and first sets of bars in middle panel of Figure 9.4.4) suggesting that this approach, which uses both space and time information, provides potentially useful estimates of attributable temperature changes from those models for which the full range of simulations are not available. Applied to the full range of models available as part of the IPCC AR4 20C3M project, this analysis supports the conclusion from more comprehensive analyses that there has been a significant contribution to 20th century warming from increased greenhouse gas concentrations with cooling from aerosols counteracting some of the greenhouse warming (Figure 9.4.4) The results also show a small warming from increasing solar forcing during the first half of the 20th century. A simpler approach is to fit 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, again, be inferred from the observed global mean time series using optimal detection (Stone et al., 2005a; Stone et al., 2005b; Stone and Allen, 2005b). Applied to the full range of models available as part of the IPCC AR4 20C3M project, results are generally consistent with space-time analyses (Figure 9.4.4) although uncertainties are greater. However when the same methodology is applied to a 62 member ensemble of simulations of the CCSM1.4 climate models, Stone et al. (2005b) found a negligible contribution to warming since 1940 from either solar or volcanic forcing. By tuning an EBM to the observations, thereby using a climate model solely to estimate internal variability, Stone and Allen (2005b) detected the effects of greenhouse gases and of tropospheric sulphate aerosols, but not the effects of volcanic stratospheric aerosols and solar irradiance changes, in the observed record between 1900 and 2004.

Gillett et al. (2002c) 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 in 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 intra-model uncertainty has the same covariance structure as internal variability. Huntingford et al. (2005) 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 (last sets of bars in Figure 9.4.4) support those results generated by spatio-temporal analyses on single models (first four sets of bars in Figure 9.4.4) in showing that greenhouse gases have likely 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 greenhouse gases likely caused more than the observed warming over the last 50 years of the 20th century, with a significant, but possibly small, net cooling from aerosols. By scaling up models whose transient response to greenhouse gas increases is lower than observed and scaling down those models whose response is higher than observed, attribution analyses bring the attributable greenhouse gas component into better agreement than the raw unscaled model results. 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,c; Stott et al., 2005a), in addition to the trend in hemispheric warming contrast itself, whose strength is model dependent (Hegerl et al., 2001)

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. As discussed by Kettleborough et al. (2005), 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 have also taken prior uncertainty and 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 trends (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 natural internal variability alone, by natural internal variability and greenhouse gas forcing (G), or by natural internal variability and the combined effect of greenhouse gas and sulphate aerosol forcing (GS). Assuming that these three possibilities were equally likely a priori, Schnur and Hasselman inferred that the odds of hypotheses G and GS against the natural-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. In a combined analysis, G and GS are still both detected, with the odds of GS over G being only slightly larger than one. Compared to conventional analyses, the inclusion of the model error structure in the Bayesian analysis leads to a downgrading of the information on the impact of sulphate forcing. They postulate that this occurs because the model uncertainty in the response to aerosols is much larger than that for greenhouse gases.

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 (2005a) find decisive evidence for an "all forcings" explanation over the 20th century, as against alternative explanations involving internal variability only, natural forcings only, and greenhouse gas increases only. They also found strong evidence for both "all forcings" and naturally forced explanations of temperature changes during the 1900–1949 period. They report that their results are insensitive to inter-model uncertainties as estimated from the IPCC AR4 20C3M simulations and to prior probabilities.

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). Gillett et al., 2002b (see also Hegerl and Allen, 2002) showed that this could be traced to differences in the response to sulphate aerosol forcing in the different climate models used. Bayesian studies find 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 (Schnur and Hasselman, 2005), although there is some evidence from these studies that the combination of increases in greenhouse gases with aerosols and natural forcing factors provides a better explanation for past temperature changes than increases in greenhouse gases alone (Smith et al., 2003).

Lee et al. (2005a) 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. (2005a) evaluate the evidence for the presence of the GS signal, estimated from two versions of the CCCma CGCM, 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 the early and latter halves of the 20th century regardless of the choice of prior distribution. On the other hand, evidence for attribution, as stringently defined 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. (2005a) 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. (2005b) have considered predictability of decadal mean temperature increments relative to the preceding three decade "normal" that arise 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. (2005a) 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 9.4.5).

[INSERT FIGURE 9.4.5 HERE]

9.4.1.6 Implications for future warming rates

Knowledge of how the observations constrain the likely contributions of greenhouse gases and other forcing factors to past temperature change (Section 9.4.1.4) also provides observational constraints on likely future rates of warming. Scaling factors derived from detection analyses determine how much simulations from comprehensive coupled atmosphere-ocean global climate models should be scaled up or down to remain consistent with past temperature change. Assuming that fractional error in model predictions of global mean temperature change is constant, these same scaling factors can be applied to predictions of future change, in a manner analogous to the Model Output Statistics (MOS) technique which has a long history of application in weather forecasting. This linear relationship between past and future fractional error in temperature change is sufficiently robust over a number of realistic forcing scenarios to introduce little additional uncertainty (Kettleborough et al., 2005). Diagnostics obtained from these approaches can be used to project uncertainty in past climate change into the future (Allen et al., 2002; Allen and Stainforth, 2002; Allen and Ingram, 2002; Stott and Kettleborough, 2002; see Section 10.5.4.5). Uncertainties calculated this way are likely to be more reliable than ranges of forecast uncertainty based on ensembles of opportunity of coupled ocean-atmosphere general circulation models, which could provide a misleading estimate of forecast uncertainty (Allen et al., 2002; Allen and Stainforth, 2002).

Several authors have used such an approach. Allen et al. (2000) obtained uncertainties in 21st century warming rates under the IS92A scenario using this approach for a range of coupled climate models. Stott and Kettleborough (2002) derived separate scaling factors on the response patterns to greenhouse gas, aerosol and natural forcings for the HadCM3 model and used them to determine probabilistic forecasts of global mean temperatures for four representative SRES emissions scenarios. Global mean temperature rise was found to be insensitive to differences in emission scenarios over the first few decades of the 21st century but large differences between emission scenarios emerge by the end of the 21st century. Stott et al. (2005a) compared observationally constrained predictions from three coupled climate models with a range of sensitivities. Results showed that predictions from more and less sensitive models are brought into better agreement, demonstrating that predictions produced in this way are relatively insensitive to the model used to produce them. The transient climate response following a 1% per year increase in carbon dioxide is likely to be greater than 2°C per century (see Section 9.6.1, Figure 9.6.1) and warming rates less than 1.5°C per century are unlikely under the SRES A2 scenario.

9.4.1.7 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., see Ramanathan et al., 2005) and modelling studies have shown that the global forcing from black carbon could be greater than 0.5 W m^{-2} (Hansen and Nazarenko, 2004; Jacobson, 2001), 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 and because the modelling uncertainty is large (see Section 9.2, Roberts and Jones, 2004). Land use changes are another forcing that could be potentially important, particularly on regional scales (Tett et al., 2005), and could explain discrepancies during the late 19th and early 20th centuries between observations and model simulations that do not account for such changes (see Section 9.3). Forcing from land use changes is expected to be negative globally (see Section 9.2.1.1) implying that the net global mean attributable warming could be slightly greater than estimated from calculations that neglect this forcing. Attribution analyses that use recent model simulations which include carbonaceous aerosols and land use changes, such as the MIROC model (fourth set of bars in Figure 9.4.4) and the HadGEM1 model (Stott et al., 2005b), 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). The implication of this study is that attribution of the net aerosol contribution to past global-mean temperature change is likely to be fairly robust to inclusion of black carbon in future analyses but that separating sulphate cooling from warming due to black carbon will require a careful choice of suitable diagnostics. However there are likely to be greater differences in the patterns of response in some regions such as India (Ramanathan et al., 2005) and 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), and there has been some recent work on incorporation of modelling uncertainty of response to a particular forcing (Huntingford et al., 2005). 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, the estimated uncertainties in reconstructions of past solar forcing have increased (see Chapter 2 and Gray et al., 2005) and many climate models used in detection studies do not include dynamical and chemical processes in the stratosphere associated with the atmosphere's response to changes in solar irradiance (Gray et al. 2005). Similarly, 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; Amman et al., 2003; Andronova et al., 1999). 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., 2005b).

Systematic biases resulting from missing physics in model simulations or imperfect understanding of the physical and chemical processes involved in climate change are not explicitly included in the uncertainties derived using detection analyses and could lead to errors in detection results. Although this type of uncertainty has not been included in detection results to date, the ability of climate models to simulate large-scale temperature changes during the 20th century (e.g., see Figure 9.4.1) and the ability of detection results to provide a consistent explanation for 20th century temperature change (e.g., Figure 9.4.4) indicate that such 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 originates with the estimates of internal variability that are required for all detection analyses. These estimates are generally model-based because the observational record is too short to provide reliable natural variability estimates on the space and time scales considered in detection studies. However, models would need to underestimate variability by large factors (over 2 for HadCM3 which has larger variability than many models, Tett et al., 2002) to nullify detection of greenhouse gases in near surface temperature data. The detection of the effects of other forcings, including aerosols, is likely to be more sensitive (an increase of 40% in the estimate of internal variability is enough to nullify detection of aerosol and natural forcings in HadCM3, Tett et al., 2002).

Few detection studies have explicitly considered the influence of observational uncertainty on near-surface temperature changes. 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 (see Chapter 3), although these errors are estimated to be relatively small and more important for the first half of the 20th century than the latter half. 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 regions such as Europe that have centuries-long instrumental records, late 20th century temperatures are the warmest observed (Chapter 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 (Figure 3.2.8, Section 3.2.2.7). 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 (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), although the processes that generate variability on small spatial scales may not be well represented in models. Thus, detection at an individual grid box does not necessarily imply that temperatures are controlled largely by external forcings and 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 (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.

One method for assessing climate change on sub-global scales was carried out by IDAG (2005). By comparing analyses of full space-time fields with results obtained after removing the linear trend in the global mean and after removing the annual global mean from each year in the analysis, they found that the detection of anthropogenic climate change is driven by the pattern of the observed warming in space and time, not just by consistent global mean temperature trends between models and observations. These results suggest that greenhouse warming should also be detectable on sub-global scales (see also Barnett et al., 1999). However, their results also show that uncertainties on the scaling factors, or equivalently the estimated contribution to 20th century warming, increases, 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. They find evidence that climates in both domains have been influenced by anthropogenic emissions during the latter half of the 20th century. This finding is robust to the exclusion of North Atlantic Oscillation (NAO/AO) related variability (Thompson and Wallace, 1998), which may itself be related to anthropogenic forcing (see Gillett et al., 2000, 2003b; Fyfe et al., 1999; Shindell et al., 1999). As the spatial scales considered become smaller, the uncertainty in estimated signal amplitudes (as demonstrated by the size of the vertical bars in Figure 9.4.6) 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 amount of estimate of internal climate variability from the control simulation is doubled.

[INSERT FIGURE 9.4.6 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 (Zhang et al., 2005). Figure 9.4.7, from IDAG (2005), shows that HadCM3 reproduces many features of the observed temperature changes and variability in the different regions. 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. (2005) 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. The recent warming observed over the Tibetan Plateau is reproduced by 20C3M simulations of the MIROC and GFDL CM2.1 models but not by their control simulations, suggesting an anthropogenic influence on increasing temperatures there (Duan et al., 2005).

These results, taken together with the sub-global detection analysis described above, provide compelling evidence of human influence on regional climates. Nevertheless, difficulties remain in simulating temperature changes in some regions of the World. 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 9.4.7). 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 has cooled over the last 100 years, suggesting that multi-decadal internal variability could be responsible (Kunkel et al., 2005). Zhou and Yu (2005) find that most IPCC AR4 20C3M simulations fail to simulate the warming observed in China in the first half of the 20th century and many have difficulty reproducing differential warming trends in North and South China in the second half of the 20th century. 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.

[INSERT FIGURE 9.4.7 HERE]

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. (2005) 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.

A very different approach to the studies described above is to compare observed temperature changes with atmosphere-only general circulation model simulations forced with observed sea surface temperatures. Sexton et al. (2003) proposed a methodology for extracting estimated signals of land surface air temperature for anthropogenic and natural effects from SST forced simulations and their interactions. 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, although the causes of changes in SSTs remain unexplained using this methodology. In agreement with previous studies using coupled models (Gillett et al., 2004c; Meehl et al., 2004; see Section 9.4.1.2); Sexton et al. (2003) found that the greenhouse gas and direct sulphate aerosol effect responses add linearly. However, Sexton et al (2003) also detected a non-linear interaction between the effects of greenhouse gases and the indirect effect of sulphate aerosols that was similar in magnitude to the individual effects of changing tropospheric and stratospheric ozone concentrations and the direct sulphate aerosols effect, although interpretation of this result is complicated by potential inconsistencies between the imposed SSTs and the model's atmospheric response to the imposed forcings.

9.4.2.3 Studies based on 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. 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; Stone and Weaver, 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 failing to capture 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 Section 3.4.3.1), a result supported by other analyses (Stone and Weaver, 2002, 2003; Dai et al., 1999).

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 (2005a) 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., 2005b). By removing the rainfall related component of Australian temperature variations, thereby enhancing the signal-to-noise ratio, Karoly and Braganza (2005b) detected an anthropogenic warming signal in south eastern Australia, although their results are affected by some uncertainty associated with their removal of rainfall related temperature variability.

9.4.3 Surface Temperature Extremes

9.4.3.1 Observed changes

Observed changes in extremes are consistent with the observed warming of the climate (Alexander et al., 2005). 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. (Section 3.8.2.1).

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. In the absence of detection results for extreme events, studies based on model output alone can be used to develop suitable approaches for early detection. Hegerl et al. (2004) investigated anthropogenic changes in daily minimum and maximum temperature over land in two atmosphere-ocean general circulation models, using indices scanning the transition from mean to extreme climate events within a year. They found a subtle but significant difference between changes in extremes and seasonal means over large areas of the globe, and estimated the signal-to-noise for changes in extreme temperature to be nearly as large as for changes in mean temperature.

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., 2005). Kiktev et al. (2003) analysed a subset of such indices, by using fingerprints from atmospheric model simulations driven by prescribed sea surface temperatures and a bootstrap method for significance testing. 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 significantly improves the simulation of these changing temperature extremes.

Christidis et al. (2005) analysed a new gridded dataset of daily temperature data (Caesar et al., 2005) using the indices shown by Hegerl et al. (2004) to have a potential for attribution. Robust anthropogenic changes were detected in indices of extremely warm nights. Human influence on cold days and nights was also detected, although less robustly, and with some indication that simulated changes were underestimated. Christidis et al. (2005) found no detection of a significant human influence on extremely warm days, 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

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.

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) but nevertheless global mean temperatures vary from year to year around the underlying forced changes due to internal variability. For regional changes in mean temperature and for temperature extremes, the chaotic component from variability becomes more important relative to the predictable changes from external forcings.

Most attribution studies have considered underlying deterministic changes rather than the actual events themselves. Allen (2003) and Stone and Allen (2005a) proposed a methodology for making quantitative attribution statements about individual 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 single event in a chaotic system 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. Changes in the probability and recurrence time of extreme rainfall, temperature and storminess events are expected under climate change conditions (e.g., Kharin and Zwiers, 2000, 2005), suggesting that this kind of quantitative risk analysis will become more important in the future.

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 throughout 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 500 years. Schär et al. (2004) showed that the central European heat wave of 2003 was 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 Thermohaline

Circulation, which could drive multi-decadal variations in western European summer climate (Sutton and Hodson, 2005). Klein Tank et al. (2005) show evidence that there have been distinctive patterns of change in European temperature variance in spring and summer that are not consistent with patterns of change in temperature variance expected from natural variability. Meteorological aspects of the summer 2003 heat wave are discussed briefly in Chapter 3, Box 3.5, and in more detail by Fink et al. (2004) and Black et al. (2004).

There could be an increase 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 (Schär et al., 2004 and Hegerl 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. They were therefore able to consider the problem as a two stage problem. By carrying out an optimal detection analysis and comparing observed and modelled European temperature changes, they first estimated the change 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 9.4.7 shows the estimated likelihood of the risk (probability) of exceedance of a 1.6°C threshold in the presence (red line) and absence (green line) of anthropogenic change, expressed both as a frequency (number of occurrences per thousand years, top axis) and as a return period (bottom axis). The clear shift from the green to the red distribution (Figure 9.4.8a) implies an FAR distribution whose mean is 0.75, corresponding to an increase in risk of a factor 4 (Figure 9.4.8b).

Although Stott et al estimated that the summer heat wave was still a relatively rare occurrence (the current probability of exceeding their chosen threshold in their model simulations was 1 in 250 years) the HadCM3 simulations showed a rapidly increasing risk such that by the 2040s half of all summers (averaged over their large region of continental Europe and the Mediterranean) would be warmer than 2003 under the SRES A2 emissions scenario. This was due in their model largely to a change in mean temperature with little change in variability over the very large region they considered. Over smaller land areas of the continent, drying of soil and vegetation feedbacks leads to an increase in variability in addition to an overall warming trend in model predictions. Every second summer in Switzerland can be expected to be hotter than 2003 by the 2071–2100 period under the same SRES A2 scenario (Schar et al., 2004; Beniston et al., 2004), and heat waves are expected to become more intense, more frequent and longer lasting both in Europe and North America according to simulations of the PCM model (Meehl and Tebaldi, 2004), although some models may over-estimate soil moisture feedbacks and so exaggerate increases in extreme temperatures (Rowell and Jones, 2005). Hegerl et al. (2004) showed that two climate models simulate a stronger change in European hot summer days in future than in summer means, widening the future temperature distribution and therefore leading to stronger warm extremes. This study also found a stronger increase in the temperature of cold winter days in future than in winter means, narrowing the future winter temperature distribution, and therefore producing less extreme cold winter temperatures.

Tebaldi et al. (2005) 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.

[INSERT FIGURE 9.4.8 HERE]

9.4.4 Free Atmosphere Temperature

9.4.4.1 Observed changes

Radiosonde and satellite temperature measurements show that the troposphere has warmed and the stratosphere has cooled since the 1960s (Section 3.4.1). Over the shorter period for which there have been both satellite and radiosonde data, there remains considerable uncertainty as to the differential warming rates seen at the surface and in the free atmosphere. Some analyses indicate a significant difference, with relative cooling aloft compared to warming at the surface. However, other analyses indicate that there have been warming rates that are consistent with or larger than those at the surface, as expected from models.

Record lengths of 25 years are now possible from satellites. At the time of the TAR only one group (University of Alabama – Huntsville UAH) had produced a 20-year climate data record from operational Microwave Sounding Unit (MSU) data. Since then other analyses of the MSU data have been produced (Mears et al., 2003; Mears and Wentz, 2005 (Remote Sensing Systems – RSS); Grody et al., 2004). Global-mean trends in the mid-troposphere (channel T2) retrieval differ greatly between the three analyses (see Section 3.4.1). If these differences are real then uncertainties due to different retrieval methods are very large (Thorne et al., 2005b). Fu et al. (2004) demonstrated that T2 temperature trends are contaminated by stratospheric cooling, and that when retrieved stratospheric temperatures (T4) are used to remove this influence, T2 temperature trends are in closer agreement with surface trends. Both UAH and RSS have also produced a lower-tropospheric retrieval (T2LT) and these differ by a similar magnitude to their T2 retrievals. Both T2LT retrievals are now internally consistent following the identification of a flaw in the diurnal correction applied by UAH (Mears and Wentz, 2005) and its subsequent removal. This flaw had led to a spurious cooling in the tropics in the UAH lower-tropospheric retrieval.

Concurrently, there have been two major efforts to create new homogenised records from radiosondes since 1958. LKS (Lanzante et al., 2003a, b) produced a dataset based on 87 globally distributed radiosonde stations, which is updated in real-time (RATPAC; Free et al., 2005). HadAT is based on a much larger network of radiosonde stations (Thorne et al., 2005a). There is evidence that pervasive cooling trends due to reducing warm biases in instruments may contaminate radiosonde records and this and other effects tend to be magnified with height, reducing confidence in the radiosonde datasets as height increases (Sherwood et al., 2005; see Section 3.4.1).

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 et al., 2000; Seidel et al., 2001). Similar increases have been inferred from three different reanalysis products (ERA-15, ERA-40 and NCAR-NCEP) and from model simulations with combined anthropogenic and natural forcing (Santer et al., 2003a, c; Santer et al., 2004; see Figure 9.4.9). 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., 2003a). However, earlier model studies have found that it is difficult to alter tropopause height through stratospheric ozone changes alone (Thuburn and Craig, 2000). Santer et al. (2003a) found that the model-predicted fingerprint in 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 9.4.9).

[INSERT FIGURE 9.4.9 HERE]

9.4.4.3 Overall atmospheric temperature change

Tett et al. (2002) compared radiosonde observations of zonal-mean changes in temperature for the 1961 to 1995 period with those simulated by HadCM3. They detected an anthropogenic response, but found that its magnitude was overestimated by the model. Their results also suggested that stratospheric variability was underestimated by the model. This study was extended by Jones et al. (2003) and by Thorne et al. (2002, 2003). Jones et al carried out a space-time analysis of temperature changes in both free-atmosphere and surface observations. They concluded that the effect of greenhouses gases and sulphate aerosols was robustly detected. In addition they could detect the influence of volcanic aerosol, and less robustly, solar irradiance changes on the temperature of the free-atmosphere/near-surface. Santer et al. (2003b) also demonstrated that PCM-simulated tropospheric temperature changes agree well with observations.

Thorne et al. (2002) carried out a sensitivity study of the processing methods used in Tett et al. (2002) and Allen and Tett (1999). They used results from both HadCM2 and HadCM3 over the same 1961–1995 period and found that the combined influence of greenhouse gases and sulphate aerosols was robustly detected, even without the inclusion of stratospheric temperature changes. When stratospheric values were more heavily weighted, best-fit simulated values were inconsistent with the radiosonde observations, due to shortcomings in the models' simulations of stratospheric variability. Thorne et al. (2003) subsequently carried out a space-time analysis. This analysis differed from that of Jones et al. (2003) in that they considered 5–10 large-area average regions rather than zonal-averages and used both HadCM2 and HadCM3. Like other studies they found robust detection of an anthropogenic impact on climate. With much less confidence they found evidence of a solar and volcanic impact on the upper-troposphere but little confidence of a detectable natural influence on other diagnostics. Both models, especially HadCM3, tended to over estimate free-atmospheric warming and they therefore were not able to unambiguously attribute recent observed tropospheric temperatures changes to any combination of external forcing influences, although they did conclude that by far the most plausible causes are anthropogenic.

Crooks (2004) detected a solar signal in atmospheric temperature changes as seen in the HadRT 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 in the free atmosphere, which was also seen by Stott et al.. (2003b) for near-surface temperatures, although 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., 2005b). 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

Radiosonde observations indicate less overall warming in the troposphere compared to the surface since 1979 (Section 3.4.1.5), in contrast to coupled model simulations, and this discrepancy is largest in the tropics (Santer et al., 2005a). However, the discrepancy between simulated and observed differential warming in the tropics is smaller based on satellite measurements of tropospheric temperature change (Santer et al., 2005b), particularly when the effect of the cooling stratosphere on tropospheric retrievals is accounted for (Section 3.4.1.5). External forcing other than greenhouse gas changes also helps to reconcile some of the differential warming, since both volcanic eruptions and stratospheric ozone depletion may have cooled the troposphere by more than the surface over the last several decades (IPCC, 2001; Santer et al., 2000; Santer et al., 2001; Free and Angell, 2002; CCSP, 2005). 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 AO/NAO, 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, 2005).

Concerns have mainly centred around the deep tropics where models exhibit a pronounced warming maximum aloft, whereas until recently the available observations suggested little if any trend. Santer et al. (2005b) undertook an intercomparison between RSS, UAH, HadAT, and RATPAC and simulations from 19 IPCC AR4 models. On monthly and annual timescales variations of temperature at the surface are amplified aloft in both models and observations by consistent amounts. It is only on longer timescales that disagreement arises. Only RSS is consistent with the models on both short and long timescales. While this does not prove that RSS is correct it does demonstrate the importance of multiple observational records, preferably from independent observational platforms.

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 record. A large number of 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. However, none of these models is able to simulate the warming observed over the last 50 years when they include just natural factors. These studies demonstrate that anthropogenic forcings dominate recent global warming. 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 greenhouse gases were responsible for more than the observed warming over the last fifty years. They show that there has been a significant cooling from aerosols counter-acting some of the greenhouse warming (Figure 9.4.4). Spatial information is evidently necessary to reliably detect the influence of aerosols. In particular aerosols are expected to cause differential warming and cooling rates between the northern and southern hemisphere, and this fingerprint helps 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. Bayesian studies, which find very strong evidence for an anthropogenic influence on climate, find weaker evidence for a significant effect of aerosols. However, clear evidence from optimal detection studies of a significant aerosol cooling over the last five decades indicates that, even in the light of continuing uncertainties in aerosol forcing and the climate's response, it is likely that greenhouse gases were responsible for 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 9.4.7) and the detection of significant anthropogenic effects on individual continents (Figure 9.4.6) provide compelling evidence for human influence on regional climates. 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 clearly identified in a variety of different studies in a number of disparate regions.

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 with a less robust detection of a human-induced warming of the coldest nights and days of the year. Many important impacts of climate 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. A calculation of the change of risk attributable to anthropogenic factors (fraction of attributable risk, FAR) has been made for the European heat wave of 2003. It was estimated that, using a threshold that was exceeded in 2003, but 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 9.4.8). 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 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 9.4.9).

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 observational dataset used in the comparison. The IPCC AR4 simulations are remarkably consistent in their predictions of tropical tropospheric changes, but only one observational estimate follows this behaviour. 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

Our 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. Since the TAR there has been an accumulation of evidence for climate change within the ocean, both at regional and global scales (Chapter 5). The overall heat content in the world ocean is estimated to have increased by 14.5×10^{22} J during the period 1955–1998 (Section 5.2.2; Fig. 5.2.1). This overall increase has been superimposed on estimated decadal variations of $\pm 5 \times 10^{22}$ J. The fact that the ocean gained heat during the latter half of the 20th century is in itself, a strong argument for a net positive radiative forcing of the climate system. If the observed warming of the atmosphere (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., 2001; Figure 5.2.1). 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 (Section 5.2.2; Fig. 5.2.1).

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 increasing trend in global ocean heat content. In contrast, changes in solar forcing can potentially explain only a small fraction of the observed increase in ocean heat content (Crowley et al., 2003)**.** It is likely that cooling from 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). 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., 2005). 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, ocean warming due to anthropogenic factors (including well mixed greenhouse gases and sulphate aerosols) are 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 increasing trend in ocean heat content is well explained by models, the decadal variability seen in Levitus et al. (2000) 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 observed variability over much of the oceans.

9.5.1.3 Sea level

We present here a brief synthesis of sea level findings that are assessed in Section 5.5.7. Model simulations with anthropogenic forcings show a significant sea-level rise of between 0.2–0.6 mm/yr from thermal expansion and 0.2–0.3 mm/yr from glacier melt on average for the 20th century, roughly consistent with observational estimates of these terms. However, models systematically underestimate the observed sea-level rise of 1.5 to 2 mm/yr. After including estimates of smaller contributions, some of the observed rise still remains unexplained. Coupled model results indicate that natural forcings and warming before the mid-19th century are significant factors in explaining the overall evolution of sea-level rise during the 20th century. Simulations including these forcings show a fairly steady rate of rise during the 20th century, rather than the acceleration implied by anthropogenic forcings alone, for which there is no observational evidence. Anthropogenic forcings are estimated to be the largest contributor to the overall sea level rise during the 20th century.

9.5.1.4 Water mass properties

The ocean heat content analyses cited above are based on 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 (Section 5.3). Sub Antarctic mode water and the sub-tropical gyres have warmed in the Indian and Pacific basins since the 1960s and there has been a freshening of the salinity minimum layer in the Atlantic, Indian and Pacific oceans as well as a freshening of surface waters in the Atlantic (Curry et al., 2003). These changes in Antarctic Intermediate Water and North Pacific Intermediate Water are consistent with a global increase in the hydrological cycle with increased precipitation at high latitudes (Wong et al., 1999). 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 (Figure 5.3.1; Figure 5.3.9; Figure 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., 2005). 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). Banks and Bindoff (2003) show that Indo-Pacific water mass changes could provide a 'fingerprint' of anthropogenic climate change.

One possible oceanic consequence of climate change is a slowing down or even halting of the thermohaline circulation. Freshening of North East Atlantic Deep Water has been observed (Curry et al., 2003; Dickson et al., 2002; Figure 5.3.15) and has been interpreted as being consistent with an enhanced hydrological cycle and a possible slowing down of the THC. 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 upward rather than downward trend in the THC. 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. The North Atlantic freshening trend in the HadCM3 simulations originates from the Arctic Ocean where there is a decrease in sea ice and an increase in river runoff. Wu et al. (2005a) show that observed increases in Arctic river flow (Peterson et al., 2002) are well simulated by HadCM3 and propose that this upward trend could be an early indicator of an anthropogenic intensification of the hydrological cycle, since this upward trend is not seen in HadCM3 simulations including just natural forcing factors. Therefore anthropogenic factors may be driving increases in river runoff into the Arctic but the relationship between this increased source of fresh water and freshening in the Labrador Sea and the link with the overturning circulation is not yet clear. Wu et al. (2005b) propose that recent freshening in the Labrador Sea could be driven by natural rather than anthropogenic forcings. Variability in the strength of the THC could also be associated with the Atlantic Multidecadal Oscillation, a multi-decadal mode of variability (Knight et al., 2005).

9.5.2 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 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 theory indicating the direction or magnitude of such changes. Risbey et al. (2002) used models and synoptic-dynamical reasoning to assess likely changes in atmospheric circulation under enhanced greenhouse conditions. They determined that it was difficult, at this time, to confidently predict the likely change in features such as jet position or strength, or in the strength and extent of the Hadley Cell. This reduces the likelihood of successful detection and attribution of changes in circulation patterns.

9.5.2.1 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 (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. Tett et al. (2005) found no significant response in the related Southern Oscillation Index to either anthropogenic or natural forcing in multi-century integrations of HadCM3. 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). A recent survey of the simulated response to CO_2 doubling in fifteen IPCC AR4 coupled climate models (Merryfield, 2005) 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 (1999a) 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 (Section 3.6.3.1). 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 is more heavily influenced by variability at decadal and longer time-scales than are traditional ENSO indices (Mantua et al., 1997; Newman et al., 2003). Recent work suggests that the PDO is 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., Newman et al., 2003; Liu et al., 2002; 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., Salinger et al., 2001; Deser et al., 2004; Vimont et al., 2002; Arblaster et al., 2002; Reason and Roualt, 2002; Yukimoto et al., 2000; Pierce et al., 2000; 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 on the PDO based on simulations of the MIROC model (Shiogama et al., 2005).

9.5.2.2 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 (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 (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., 2002a; 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% significance 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 in response to increased concentrations of greenhouse gases (Fyfe et al., 1999; Paeth et al., 1999; Shindell et al., 1999; Gillett et al., 2003a, b; Osborn, 2004; Rauthe et al., 2004), although the simulated trend is generally smaller than that observed (Gillett et al., 2002a; 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., 2002a; Osborn, 2004; Rauthe et al., 2004). Some studies have postulated an influence of ozone depletion (Volodin and Galin, 1999; Shindell et al., 2001b), changes in solar irradiance (Shindell et al., 2001b), and volcanic eruptions (Kirchner et al., 1999; Shindell et al., 2001b; Stenchikov et al., 2005), and on the Northern Annular Mode. Stenchikov et al. (2005) examined changes in sea level pressure following nine volcanic eruptions in 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., 2001b; 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 (Miller et al., 2005; Gillett, 2005; 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 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 (Hurrell et al., 2005; Hoerling et al., 2001), 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). 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).

9.5.2.3 Southern Annular Mode

The Southern Annular Mode (SAM) is more zonally-symmetric than its Northern Hemisphere counterpart (Thompson and Wallace, 2000) (Section 3.6.5). It too has exhibited a pronounced upward trend over the past thirty 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 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 recent peak is unprecedented (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 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., 2005), 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; Stone et al., 2001; Kushner et al., 2001; Cai et al., 2003; Marshall et al., 2004; Shindell and Schmidt, 2004; Arblaster and Meehl, 2005; Stone and Fyfe, 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}, cooling of the Antarctic interior, and part of the warming of the Antarctic Peninsula (Thompson and Solomon, 2002; Carril et al., 2005), though other factors are also likely to have contributed to this warming (Vaughan et al., 2001).

9.5.2.4 Sea level pressure detection and attribution

Trends in sea level pressure changes 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]

9.5.2.5 Monsoon circulation

The current understanding of climate change in the monsoon regions remains one of considerable uncertainty with respect to circulation and precipitation (Sections 3.7.1 and 9.5.3.2). The Asian monsoon circulation in the IPCC AR4 models was found to decrease by 15% by the late 21st century, according to the ensemble mean of the IPCC model simulations (Tanaka et al., 2005; Ueda et al., 2005), 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 (South-eastern Asia, western Africa, eastern Africa, and the Australia/Maritime Continent) for the period 1950–1998. In each region they found a consistent picture of significantly diminished monsoonal circulation, although this result is uncertain due to changes in the observing system affecting the NCEP reanalysis (Section 3.7.1). So the results of Chase et al are not inconsistent with the expectation (Tanaka et al., 2005; Ramanathan et al., 2005) of weakening monsoons due to anthropogenic factors, but further model and empirical studies are required to confirm this.

9.5.2.6 Tropical cyclones

The active North Atlantic hurricane season of 2004, and the unusual development of a large cyclonic system in the subtropical South Atlantic that hit the coast of southern Brazil in March 2004 (e.g., Pezza and Simmonds, 2005), led to 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 Box 3.4, which discusses processes affecting tropical cyclone formation, frequency, and intensity, and Trenberth, 2005). Recent modelling studies continue to support this conclusion. Knutson et al. (2001) suggest that maximum surface wind speeds may increase by about 5% under CO_2 doubling, although Michaels et al. (2005) suggest that this may be an over-estimate. Knutson and Tuleya (2004) demonstrate that this result is robust across a range of choices for the CGCM that is used to set the environment for hurricane simulations and the convection parameterization used in those simulations. However, Sugi et al. (2002), in a time slice experiment with a T106 atmospheric model, suggest that tropical cyclone frequency may decrease in response to future greenhouse gas forcing. McDonald et al. (2005) and Oouchi et al. (2005) obtain similar results in time slice experiments with high resolution atmospheric models with prescribed SST and anthropogenic forcing changes. Wu and Wang (2004) examine the likely changes in large-scale environmental flows and suggest that western North Pacific tropical cyclone tracks may shift slightly south-westward over the next three decades. A fine-resolution model study by Walsh et al. (2004) suggested that the numbers of tropical cyclones near Australia would not change greatly under enhanced greenhouse conditions, but that the number of intense systems may increase, particularly at higher latitudes.

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). Bister and Emanuel (2002) found generally positive trends in tropical cyclone potential intensity (a measure of the maximum intensity expected of tropical systems) based on reanalysis data, but Free et al. (2004), using radiosonde data from tropical islands, found no consistent trend from 1975 to 1995 in either the Pacific nor the Caribbean. Trenberth (2005) suggests that there has been some increase in heavy rainfall associated with hurricanes affecting the southeast United States, due to changes in the environmental conditions in which hurricanes are embedded. Emanuel (2005) reports a marked increase since the mid-1970s of an index of the destructiveness 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 and the NAO as well as the strong warming since the mid-1970s, suggesting that the increased destructiveness 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 the small relative increase in intensity expected from a doubling of CO_2 (Pielke et al., 2005), and the lack of a strong consistent signal in the changes in cyclone numbers in model experiments under enhanced greenhouse conditions, (Knutson and Tuleya, 2004; Walsh, 2004).

9.5.2.7 Extra-tropical cyclones

Simulations of the 20th century in the IPCC coupled 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, 2005). 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 (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. However, the reanalyses from which the observed changes have been deduced have problems due to changes in observing systems, thus we cannot yet conclude that an anthropogenic effect on mid-latitude storm tracks has been detected.

9.5.3 Precipitation

9.5.3.1 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. Radiosonde observations of water vapour over Northern Hemisphere land generally indicate increasing trends in specific humidity (Section 3.4.2.1). 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 global oceans (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 both 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, indicating that the water vapour feedback is realistically simulated.

[INSERT FIGURE 9.5.3 HERE]

9.5.3.2 Global precipitation changes

The increased atmospheric moisture content associated with warming might be expected to lead to increased global mean precipitation. However, precipitation is also strongly influenced by changes in the tropospheric energy budget and the atmospheric circulation, so spatio-temporal patterns of precipitation change are likely to be complex. Global terrestrial annual mean precipitation showed a small upward trend over the 20th century of approximately 2.1 mm/decade based on GHCN data (Section 3.3.2.1). However, the record is characterised by large interdecadal variability, and global terrestrial annual mean precipitation based on the New et al. (2000) record shows almost no trend since 1940 (Figure 9.5.4).

9.5.3.2.1 Detection of external influence on precipitation

Several studies have demonstrated that simulated terrestrial precipitation in climate model integrations including both natural and anthropogenic forcings is significantly correlated with that observed (Allen and Ingram, 2002; Lambert et al., 2004; Gillett et al., 2004b), thereby detecting external influence in observations of precipitation. Lambert et al. (2005) examine simulated precipitation changes in simulations of nine IPCC AR4 coupled models including anthropogenic and natural forcing (Figure 9.5.4). 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 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 therefore suggest that natural forcings such as volcanic aerosol are likely to have had a larger influence on precipitation than have greenhouse gas changes thus far, and this is consistent with simulations of the response to volcanic forcing (Robock and Liu, 1994; Broccoli et al., 2003). Simulations and observations indicate that global precipitation decreases by approximately 1 to 3% following large volcanic eruptions (Robock and Liu, 1994; Broccoli et al., 2003). 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]

Mitchell et al. (1987) argue that precipitation changes are 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 greenhouse gas increases (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 explains why precipitation responds 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. It also explains why anthropogenic influence is not yet detectable in measurements of global mean precipitation (Gillett et al., 2004b; Ziegler et al., 2003). 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.

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) were insufficient to explain the increased run-off.

Burke and Brown (2005), 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 simulate the observed increases in the global frequency of droughts in the second half of the 20th century, although the model trend was weaker than that observed. 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 Brasil and northwest Australia). While this was not a formal detection study, it does suggest that the observed global trend towards increased droughtiness in the second half of the 20th century is at least partly due to external forcings, although the relative contributions of natural external forcings and anthropogenic forcings has not been assessed.

9.5.3.2.2 Changes in extreme precipitation

Climatological data show that the most intense precipitation occurs in warm regions (Easterling et al., 2000). Also, 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 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. (2005) point out that since the amount of moisture in the atmosphere is likely to rise much faster (as a consequence of rising temperatures) than the total precipitation, this should lead to an increase in the intensity of storms, offset by decreases in duration or frequency of events.

Model results also suggest that future changes in precipitation extremes will likely be greater than changes in mean precipitation (see, for example, Meehl et al., 2000; Kharin and Zwiers, 2000, 2005; Kharin et al., 2005; Semenov and Bengtsson, 2002; Groisman et al., 2005; Pall et al., 2005; Emori and Brown, 2005). Simulated changes in globally averaged annual mean and extreme precipitation appear to be quite consistent between models. Allen and Ingram (2002) suggest that while mean precipitation is constrained by the energy budget of the troposphere, extreme precipitation is constrained by its 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 (Pall et al., 2005). 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. 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., 2005) and the analysis method that is employed (IPCC, 2001; 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. (2005) reported that eight GCMs run for the IPCC AR4 showed general agreement of a trend towards greater frequency of heavy-precipitation events over the past four decades, most coherently in the high latitudes of the Northern hemisphere.

9.5.3.3 Regional precipitation changes

Regional trends in precipitation are likely to exhibit strong spatial variations, even if global precipitation increases somewhat as a result of warming, because of the dependence of precipitation on atmospheric circulation. Trends in observed annual precipitation during the period 1901 to 2003 are shown in Figure 3.3.2 for regions in which data is available. Annual mean precipitation increased over most of North and South America, although negative trends were observed in some regions such as the south-western United States and parts of the western coast of the South America (Marengo et al., 2004). Over Europe, precipitation has increased in the north and decreased in the south, and over Australia it has increased over much of the continent but decreased in the southwest. Very large negative trends in regional annual precipitation occurred over Africa, particularly over the Sahel region. Although anthropogenic influence has not been detected in spatial patterns of precipitation change, Milly et al. (2005) demonstrated that observed 20th century changes in runoff are significantly correlated with those simulated in a subset of nine of the IPCC 20th century simulations. However, runoff also depends on evaporation, thus this result does not reflect changes in precipitation alone but also may reflect changes in transpiration by plants in response to changes in CO_2 (Gedney et al., 2005).

9.5.3.3.1 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). There has been a partial recovery since about 1990, although rainfall has not returned to levels typical of the period 1920–1965. Three main hypotheses have been proposed as a cause of the extended drought (Zeng, 2003): 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). Taylor et al. (2002) examined the impact of land use change with a GCM forced only by estimates of 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 (Giannini, 2003; Rowell, 2003; Bader and Latif, 2003; Held et al., 2005; Haarsma et al., 2005; Hoerling et al., 2005a; see also Figure 9.5.5) , consistent with earlier findings (Folland et al., 1986). These studies differ somewhat in terms of which ocean SSTs they find to be most important: Gianni et 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 et al. (2003) ascribe decadal variations in Sahel rainfall largely to variations in Mediterranean SST. 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 sea surface temperatures are likely to have been the dominant influence on rainfall in the Sahel, rather than land use changes (Taylor et al., 2002). But what has caused the 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. The results to date 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 GFDL-CM2.0 and CM2.1 models both 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 ensemble mean does not reproduce the magnitude of the observed drought in the 1980s, but one of the ensemble members does approximately do so. The spatial pattern of the trends in rainfall also shows some agreement with observations. However, their results are in contrast to most of the IPCC AR4 20C3M simulations, which generally show little change or a small increase in future Sahel rainfall in response to anthropogenic forcing (Haarsma et al., 2005).

9.5.3.3.2 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) conclude that land cover change could not be the primary cause of the rainfall decrease, and re-affirm the conclusion of IOCI (2002) that both natural variability and the enhanced greenhouse effect probably contributed to the rainfall decrease.

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 9.5.2.3). 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. Timbal et al. (2005) detect an anthropogenic influence in SW Australian precipitation, using climate change signals that were downscaled from the PCM. However, such local detection results should be treated with caution in the absence of evidence of an anthropogenic influence on precipitation on global scales where, presumably, signal to noise ratios would be larger.

9.5.3.3.3 Monsoon precipitation

Observations show that overall South Asian, North African and North American monsoon precipitation decreased somewhat in the mid-1970s (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 9.5.2.5). However, by simulating 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 simulated in simulations with greenhouse gas and sulphate aerosol changes only.

9.5.4 Cryosphere Changes

Widespread warming would, in the absence of other countervailing effects, be expected to lead to declines in sea ice, snow, and glacier and ice-sheet extent and thickness.

9.5.4.1 Sea ice

The annual mean area of Arctic sea ice has decreased by about 2.4% per decade over the past quarter century, with stronger declines in summertime than winter (Johannessen et al., 2004; Chapter 4). There has also been some thinning of the ice, although the amount of thinning is still in dispute (Johannessen et al., 2004). 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 decline observed. 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”.

Gregory et al. (2002b) also found a decline in Antarctic sea ice extent in their model, whereas there is no record of such a decline in the observations since at least 1973. They suggested that the lack of consistency between the observed and modelled decline might reflect an unrealistic simulation of regional warming around Antarctica, rather than a deficiency in the ice model. 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 19792000, more consistent with the observations of temperature and the lack of a decline in sea ice extent.

9.5.4.2 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 (Diaz et al., 2003; Figure 4.2.1). The same period and season has exhibited a large decline in the area above the freezing level, and all major mountain chains exhibit upward shifts in the height of the freezing level surface. The most likely explanation for the decline in snow cover is that warming has increased snow melt, or reduced the proportion of precipitation falling as snow.

Mote (2003) demonstrated that snow mass in the Pacific Northwest region of North America had declined substantially at most locations, especially below the 1800m elevation level, and that these declines coincided with significant increases in temperature and occurred irrespective of whether precipitation declined or increased. Mote concluded that positive snow-albedo feedback was likely to be contributing to springtime warming and decline in snow cover, both at low elevations and in the mountains. Stewart et al. (2004) demonstrated that spring snowmelt in western North America had shifted earlier in the year by typically 10-20 days since the mid-20th century, and that this was due to warmer spring air temperatures. Mote et al.

(2005) extended the Mote (2003) study and demonstrated that the decline in snow mass since the mid-20th century had occurred over nearly all of western North America, especially at moderate elevations. At cold, high elevations where there has been an increase in precipitation, snow cover increased.

Nicholls (2005) demonstrated that springtime (first observation in October) snow depth in the Snowy Mountains of south-eastern Australia had declined by about 40% since 1962, while winter maximum snow depth had declined only weakly. The stronger decrease in spring snow depth is largely attributable to warming during July-September. Only a weak decline in total precipitation has occurred in this region, far too small to account for the decline in snow depth, whereas mean maximum temperature (which is strongly correlated with October snow depth, $r = -0.81$, $n = 41$) has increased about 1°C, sufficient to account for the decline in springtime snow depth.

Scherrer et al. (2004) show that decreases in late 20th century low altitude Swiss Alpine snow days are associated with increases in temperature, associated partly with NAO variability.

9.5.4.3 Alpine and polar glaciers

Meier et al. (2003) synthesised data on changes in alpine glaciers and concluded that glacier wastage has been pervasive during the last century, with small glaciers and those in marginal environments disappearing and large mid-latitude glaciers shrinking slightly. The average global mass balance has been continuously negative since 1961 with an estimated mass loss of 90 km^3/yr in water equivalent. There is evidence that glacier retreat is stronger than has been assumed previously (Paul et al., 2004) and it has been suggested that positive feedbacks associated with a decrease in glacier albedo may be accelerating glacier melt. The global shrinkage appears to imply general warming, and since glacier retreat on the century time scale is rather uniform across the globe, this suggests that the warming is a pervasive cause (Oerlemans, 2005; Chapter 4) although in the tropics changes in atmospheric moisture might be contributing (Chapter 4). 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). Balances for glaciers in north and central Europe are only weakly related to global temperatures but are strongly correlated with the Arctic Oscillation. Reichert et al. (2002) forced a glacier mass balance model for Nigardsbreen and Rhone glacier with downscaled data from an OAGCM control simulation and concluded that the rate of glacier advance during the little ice age can be explained by internal climate variability for both glaciers, but that the recent retreat cannot. This suggests that the recent retreat of both glaciers is a sign of externally forced climate change. As well, at least some polar glaciers appear to be thinning and accelerating, apparently as a result of ocean warming (Thomas et al., 2004). The mean rate of glacier retreat in the Antarctic Peninsula has been increasing since 1945 (Cook et al., 2005). The pattern of retreat is broadly compatible with retreat driven by warming, but the speed of the retreat suggests that other factors may be contributing.

9.5.4.4 Polar ice sheets

Abdalati and Steffen (2001) found that the melt extent of the Greenland ice sheet had increased nearly 1%/yr over the period 1979–1999. The correlation between melt extent and temperature was estimated to be 0.82, suggesting that the warming is a likely explanation for the increased melt extent. Melt extent in 2002 set a new record since observations first became available in 1979 (ACIA, 2004). The Greenland ice sheet is losing mass by near-coastal thinning and the West Antarctic ice sheet is probably thinning overall, despite thickening in parts (Rignot and Thomas, 2002). Increased snowfall (which would lead to thickening of the ice sheets) appears to have been more than offset by increased ice-surface melt water loss (Greenland), and increased ice flow from melt water lubrication (Greenland) (Chapter 4). Domack et al. (2005) inferred from paleo data that the 2002 collapse of the Larsen ice shelf off the Antarctic Peninsula was unprecedented in the past 11,000 years and suggest that recent warming contributed to the collapse.

9.5.5 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 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 significant 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 which 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. As yet there is no compelling evidence for a detectable anthropogenic influence 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 possibly 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 explains why human influence is not yet detectable in global precipitation, but the influence of volcanic aerosols is. However, model simulations indicate that the rainfall in the heaviest events is likely to increase in line with atmospheric water vapour concentration, and there is some observational evidence that the intensity of the heaviest rainfall events has increased, although uncertainties remain. Many models capture the observed decrease in Sahel rainfall when constrained by observed sea surface temperatures, although these sea surface temperature changes have not been clearly attributed to human influence. 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 natural variability, but consistent with the simulated response to human influence, although Southern Hemisphere sea ice extent has not declined as predicted. There is evidence of a decreasing trend in global snow cover, and widespread melting of glaciers, consistent with a widespread warming.

9.6 Observational Constraints on Climate Sensitivity

In this section we briefly review recent research to infer climate sensitivity (defined as the equilibrium global mean temperature response to a doubling of CO_2 from preindustrial levels) from observed climate changes. Such inferences, which can be made in a number of ways using instrumental or paleo data, complement other approaches to the assessment of climate sensitivity based on modelling and other types of studies. An overall summary assessment of equilibrium sensitivity and transient climate response, taking all of these approaches into account, is given in Chapter 10, Box 10.2 Precise definitions of climate sensitivity are given in the Glossary and Section 8.6.2.1. That section also cautions that climate sensitivity will depend to some extent on the mean climate state and the nature of the forcing. For example, climate sensitivity to the large negative forcing in the Last Glacial Maximum relative to the preindustrial climate (Section 9.2.1.3) may not be exactly the same as that for a positive CO_2 forcing of the same magnitude (Chapter 6). Moreover, global sensitivities do not provide information about regional climate responses and how they might vary in response to different forcings (Chapter 6) Here, we focus on what inferences can be made about climate sensitivity to increases in CO_2 based on the ability of models to simulate past climate changes, and do not attempt to specify regional sensitivity and sensitivity to forcings other than greenhouse gases.

Three basic approaches are used to infer information about climate sensitivity from observed climate changes. One approach, shortly to be introduced, identifies probabilistic ranges of combinations of climate parameters in simplified climate models for which simulations of historical climate change are consistent with changes observed during the historical period considered (Section 9.6.1). Several studies of this type have used observed surface temperature changes over the last 150 years (see Chapter 3), the estimated ocean heat uptake since 1955 based on Levitus et al. (2000, 2005), and changes in atmospheric temperatures (Forest et al., 2002, 2005, Lindzen and Giannitsis, 2002). Note that studies using radiosonde data may be affected to some extent by recently discovered inhomogeneities in radiosonde data (see Section 3.4.1.1; Forest et al., 2002, 2005 use radiosonde data in addition to surface and ocean data). One recent study also uses ERBE Earth radiation budget data (Forster and Gregory, 2005). Such studies have used either spatial patterns of temperature change, or simple temperature indices such as the global mean surface temperature or the differential warming between the SH and the NH. In addition, studies using this approach have also been conducted with paleo reconstructions of NH temperature variations of the past millennium. Note that a further variant of this approach, which derives such distributions by varying uncertain parameters in coupled climate models, is discussed in Section 10.5.4.5.

A summary of key results obtained from studies that systematically vary parameters in simplified models is given in Table 9.2.1. The table also details the methods used and uncertainties that are taken into account. Methods that incorporate a more comprehensive treatment of uncertainty generally produce wider uncertainty ranges on the inferred climate parameters. Methods that treat uncertain parameters, such as ocean diffusivity, as fixed 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.

The two other approaches that will be discussed in this section further augment the information on climate sensitivity that is available. The first of these is to infer information on the climate sensitivity from specific, well observed events in recent climate history, such as the eruption of Mt. Pinatubo in 1992 (Section 9.6.1.2). The second is to deduce information on climate sensitivity, with the aid of models, from more extended, earlier, events in climate history such as the Maunder Minimum (Section 9.6.2.2) or the Last Glacial Maximum (Section 9.6.2.3).

9.6.1 Estimates of Climate Sensitivity Based on Instrumental Observations

Detection and attribution studies rigorously compare observations with model-simulated climate change in response to external forcing. Similar methods can be employed to identify which of a large number of models with varying uncertain parameters or different external forcing, best simulate the observed change. Results from such comparisons can be used to learn about important, but unknown, parameters of the climate system. Three key parameters are (i) the equilibrium climate sensitivity α, which is the global mean equilibrium near surface temperature change in response to CO_2 doubling relative to preindustrial levels, (ii) the transient climate response, which is the global mean temperature change at the time of CO_2 doubling in a scenario where CO2 increases by 1%/yr, and (iii) the rate of ocean heat uptake. These studies can also be used to make inferences about the magnitude of total external forcing or its individual components, particularly aerosol forcing, as discussed in Section 9.2.

The basic approach that is used is to vary the parameters of interest within plausible ranges, and then to quantify the similarity between model simulations and observations for each choice of parameter settings. The idea that underlies this approach is that the plausibility of a given combination of parameter settings can be determined from the agreement between the resulting simulation of historical climate and observations. This is typically evaluated by means of Bayesian methods (see Appendix 9.B for methods). Such studies require very large ensembles of simulations of historical climate change (ranging from several hundreds to thousands members). Parameters such as the climate sensitivity can only be varied directly in simple climate models or earth system models of intermediate complexity (the so-called EMICS; see Chapter 8) and thus most studies to date have been performed with such models.

The approach is essentially Bayesian because the sampling that is performed on the parameters of interest in effect specifies a prior distribution on those parameters. Frame et al. (2005) point out that the results of such studies are sensitive to the choice of prior distribution. Prior assumptions that assume that different equilibrium climate sensitivities are equally likely (“uninformed” or “flat” prior) will use a different strategy to sample the parameter space than studies in which different feedback values (i.e. the inverse of climate sensitivity) are assumed to be equally likely, or in which different transient climate responses are assumed to be equally likely. Since the observational constraints are still weak (as will be shown below), these prior assumptions matter and it is therefore important to define a sampling strategy that is suitable to the problem to be considered, for example, sample a 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.

9.6.1.1 Estimates of climate sensitivity based on 20th century warming

Several studies have used the approach described above to infer probability distributions for the equilibrium climate sensitivity or for the transient climate response from the observed warming of the last 150 years. Results and assumptions are listed in Table 9.2.1, a comparison of some probability density functions for equilibrium climate sensitivity is shown in Figure 9.6.1, and some estimates of transient climate response are displayed in Figure 9.6.2.

Wigley et al. (1997) laid the groundwork by simulating global mean temperature change over the instrumental period using a range of climate sensitivities in a simple climate model. They concluded that uncertainties made it impossible to use observed global temperature changes to constrain climate sensitivity beyond the range explored by climate models (1.5 to 4.5°C), and that the upper end of the range was particularly difficult to constrain, a conclusion confirmed by subsequent studies. Similarly, Tol and De Vos (1998) find poorly constrained upper limits for the equilibrium climate sensitivity in a Bayesian approach, testing a variety of prior assumptions using a simple statistical climate model.

Andronova and Schlesinger (2001) used a simple climate model consisting of a hemispheric Energy-Balance-Model (EBM) coupled to a simple ocean model to simulate the change in NH and SH annual-mean temperatures from 1765 to 1997. Radiative forcing was varied, including the effects of five forcing factors (greenhouse gases, tropospheric ozone, anthropogenic sulphate aerosols, solar and volcanic) in sixteen combinations. The climate sensitivity and unknown sulphate forcing where varied systematically for each combination of forcing factors. Considering the resulting collection of likelihood functions, they estimated that there was a greater than 50% likelihood that climate sensitivity lays outside the 1.5 to 4.5°C range.

Knutti et al. (2002) systematically varied the climate sensitivity, ocean heat uptake and ocean mixing parameters in an EMIC over plausible ranges. External forcing uncertainty was represented by interpreting the IPCC (2001) forcing uncertainty ranges as four standard deviation confidence intervals. Their simulations were evaluated against observed 20th century ocean and surface air temperatures. Using the time evolution of global surface warming and ocean heat uptake as constraints, and taking uncertainties into account, Knutti et al. (2002) found no reliable upper limit on climate sensitivity. They also found that the available data on ocean heat uptake does not constrain their ocean mixing parameter. However, they determined that the indirect aerosol forcing and total aerosol forcing are both very likely (95%) negative, but 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 (Section 9.2.1.2, Table 9.2.1).

Gregory et al. (2002a) constrain the climate sensitivity by considering the change in the Earth's energy balance between 1861–1900 and 1957–1994. Their calculation uses the definition of effective climate sensitivity (Murphy, 1995) in assuming that the climate system responds to warming by increasing its radiation to space proportionately to the temperature change. Energy balance requires this heat flux to equal the difference between the radiative forcing and the rate of ocean heat uptake. Best estimates of temperature change, forcing change and ocean heat content change between the two periods and their uncertainties (which are assumed to be Gaussian) are used to estimate the climate feedback parameter and its uncertainty. Since uncertainties in forcing dominate, the inverse of climate sensitivity has approximately a Gaussian distribution. Consequently, consistent with other studies, the derived likelihood function for the climate sensitivity is skewed, with the lower 5% boundary at 1.6°C and a very long upper tail (Figure 9.6.1).

Forest et al. (2002) use spatio-temporal patterns of climate change rather than using large scale indices, and. assessed the similarity between simulations and observations by means of optimal detection (Allen and Tett, 1999). They repeatedly calculated the time-space response to anomalous greenhouse gas, sulphate aerosol, and stratospheric ozone forcing using the MIT EMIC (Table 8.8.2) and compared against upper-air, surface and deep-ocean temperature observations. The climate sensitivity, net aerosol forcing and ocean heat uptake used in their simulations were selected from uniform prior distributions (i.e., all parameter combinations within pre-specified ranges were considered to be equally likely within the range explored, such as up to a sensitivity of 10°C). They estimated a 5–95% confidence interval of 1.4–7.7°C for the climate sensitivity, and a 30% probability that sensitivity was outside the 1.5 to 4.5°C range.

Forest et al. (2005) updated this study to include solar irradiance, volcanic aerosols, and land-surface vegetation forcing. They also used a larger range of AOGCM control simulations to estimate the internal climate variability. This resulted in wider uncertainty ranges on the climate parameters of interest, with 5–95% range of 2.4 to 9.2°C for climate sensitivity (Table 9.2.1). They obtained generally narrower uncertainty ranges when using expert prior distributions rather than uniform, uninformative, priors (see discussion in Appendix 9.B). The increased upper bound on their sensitivity estimate compared to the result explaining the late 20th century by anthropogenic forcing alone is due to a net cooling from volcanic aerosols, which reduces the total forcing and favours a combination of slightly higher climate sensitivity and weaker tropospheric aerosol cooling (Figure 9.6.1).

Frame et al. (2005) point out that inferences on the climate sensitivity are strongly affected by prior assumptions and the way in which uncertainty on unknown parameters is sampled. They point out that the decision as to whether to sample an unobservable parameter such as the climate sensitivity (or its inverse, the climate feedback) from a subjectively chosen prior distribution, or to sample a parameter that can be constrained by observations such as the transient climate sensitivity, depends in part upon the objective of the analysis. Sampling uniformly in climate sensitivity, they infer a 5–95% uncertainty range for the sensitivity of 1.2 to 11.8°C, which is consistent with others who take this approach (Table 9.2.1). On the other hand, sampling uniformly in transient climate response, an approach that would be appropriate if the goal is to constrain projections of future change, results in an estimated climate sensitivity of 1.2 to 5.2°C with a median value of 2.3°C. Note that this is based on an energy balance model and a multi-fingerprint separation of the observed response into greenhouse gas, aerosol and natural forcings. The latter avoids sulphate aerosol forcing uncertainty that affects parameter estimates, but adds uncertainty associated with separating the model's responses the various forcings (see discussion in Section 9.2.3).

Forster and Gregory (2005) estimate climate sensitivity based on radiation budget data from the Earth Radiation Budget Experiment (ERBE). They find a climate feedback parameter of 2.3 ± 1.4 W m^{2} K^{-1}, which corresponds to a 5–95% equilibrium climate sensitivity range of 1.0 to 4.1°C, assuming a flat prior probability in feedback parameters (note that the results would be wider with a flat prior in sensitivity or transient climate response, see Frame et al., 2005).

[INSERT FIGURE 9.6.1 HERE]

Lindzen and Giannitsis (2002) consider the difference in warming rates between tropospheric (850–300 hPa) temperatures estimated from radiosondes and surface temperatures. They note that tropospheric temperatures increased rapidly around 1976, while the surface temperature increased more slowly. Using a simple climate model, they find that if the observed surface temperature increase is interpreted as a delayed response to the tropospheric temperature increase, it is best modelled with a climate sensitivity of less than 1°C. The plausibility of this finding is limited by the lack of an explanation for the tropospheric temperature increase particularly, and neglect of the effects of external forcing. The finding in Lindzen and Giannitsis is in contrast to that of Forest et al. (2002, 2005) who also considered the joint evolution of surface and upper air temperatures.

In summary, the climate sensitivity that is inferred from constraints provided by historical instrumental data yields best estimates (mode of the estimated probability density functions) that range between 1.2 and 4°C, in agreement with other estimates derived from comprehensive climate models, and suggests a lower 5% limit on climate sensitivity of 1°C or above. The upper 95% limit 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 exceeds 4.5°C.

9.6.1.2 Estimates based on individual volcanic eruptions

Some recent analyses have attempted to constrain 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 the study of physical mechanisms and feedbacks as was done by Soden et al. (2002; see,

Figure 8.6.2) who demonstrated good agreement between simulated and observed responses, unless the (positive) water vapour feedback in their model, which had a climate sensitivity of 3.0°C, is switched off. Yokohata et al. (2005) find that a version of the MIROC climate model with a sensitivity of 4.0°C yields a much better simulation of that eruption than a model version with sensitivity 6.3°C, and determine that the cloud feedback in the latter model appears inconsistent with data. 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. 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. Both Robock et al. (2005) and Wigley et al. (2005b) question the analysis method of Douglass and Knox.

In studies based on the temperature response to single volcanic eruptions it is generally found (Wigley et al., 2005a, Frame et al., 2005) that a reliable upper limit on sensitivity cannot be established, particularly if forcing uncertainty is considered. The reason is that 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. Such differences in response are small compared to internal climate variability and the uncertainty in the rate of heat taken up by the ocean in response to a short, strong forcing.

9.6.1.3 Transient vs. equilibrium climate sensitivity.

While the equilibrium climate sensitivity determines 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 Section 8.6.2.1). The transient climate response is indicative of the temperature trend associated with external forcing, and as such it is well constrained by an observable quantity, the observed warming trend. However, the transient climate response does not scale linearly with sensitivity, particularly given uncertainty in ocean heat uptake (see discussion in Frame et al 2005), which makes it difficult to rule out high values of equilibrium sensitivity on the basis of a well constrained 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 climate sensitivity combined with small positive forcing can remain reasonably consistent with 20th century observations.

Frame and Allen (2005) point out that transient climate response may be more relevant to determining the maximum temperature change that might result under greenhouse gas forcing scenarios where concentrations first peak and then reduce rather than equilibrate. Stott et al. (2005c) 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 3°C, with a 5–95% range of f 2.2 to 4°C. We conclude that transient climate response is very unlikely to be more than 4.5°C/century in response to a 1%/yr increase in CO_2. 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.

[INSERT FIGURE 9.6.2 HERE]

9.6.2 Estimates of Climate Sensitivity Based on Paleoclimatic Data

The paleoclimate record offers a range of opportunities to try to assess the ability of climate models to realistically respond to changes in external forcing. Climate feedbacks may be different for different climatic background states, and for different seasonal characteristics of forcing. 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 in these cases, sensitivity to forcing cannot be directly compared to that to a doubling of CO_2. For climate at the Last Glacial Maximum, the sensitivity to forcing may be more comparable. In both cases, if climate models attempt to simulate past climatic conditions, model physics and the realism of the feedbacks can be validated. In some cases, the agreement between past climate and simulations with model versions with varying sensitivity to CO_2 doubling has been quantified in order to derive a probability density function for equilibrium climate sensitivity.

We discuss estimates obtained from both the paleoclimatic record of the last millennium, and from the near-equilibrium climate state of the Last Glacial Maximum (LGM). The latter give a perspective on feedbacks for a very different climate than that expected to be associated with greenhouse warming, and provides a test bed for the physics of feedbacks in climate models.

9.6.2.1 Estimates of climate sensitivity based on proxy data for the last millennium

More than one approach can be used to estimate climate sensitivity from reconstructions of the last millennium. The approach discussed here uses results on the attribution of interdecadal climate changes to external forcing (see Section 9.3.3). Another option, which will be discussed in Section 9.6.2.2, is to consider differences between different climate states that occurred during the last millennium, for example, the late Maunder Minimum relative to the present.

Hegerl et al. (2005b) use several decadal or decadally smoothed 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 Hegerl et al., 2005b) to constrain the climate sensitivity during the pre-industrial period up to 1850. As with the studies discussed in Section 9.6.1, a very large ensemble of simulations of the last millennium is performed with an energy balance model. The EBM is forced with volcanic (Crowley, 2000, updated) and solar forcing (Lean et al., 2002) reconstructions, and with changes in CO_2 forcing (see Section 9.3.3 for results on the detection of these external influences). The calculation of the probability distribution for equilibrium climate sensitivity incorporates uncertainty in the amplitude of volcanic and solar forcing. Reconstruction uncertainty is accounted for by fully incorporating uncertainty in the amplitude of one particular reconstruction, and by using several different reconstructions as possible realizations of observed climate change. Probability density functions for climate sensitivity from all reconstructions combined yield a median sensitivity of 3.4°C and a 5–95% range of 1.1 to 8.6°C, while those for the reconstruction judged to allow for the most complete assessment of uncertainty, yields a range of 1.4 to 6.1°C (see Figure 9.6.1). Probability density functions from the other reconstructions, individually, yield peak probabilities (modes) from 1.3 to 3.6°C. Generally, reconstructions with a higher amplitude of past climate variations (such as Esper et al., 2002 or Hegerl et al., 2005b) 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 statistics of volcanism that lead to a highly significant detection of volcanic response (see Section 9.3.3) in all records used in the study.

Constraints on the climate sensitivity can be further tightened by combining information from the instrumental and paleo records of the last millennium. In particular, instrumental temperature change during the second half of the 20th century is independent of the paleo record and of the instrumental data from the first half of the 20th century that is typically used to calibrate the paleo record. Thus, detection analyses on late 20th century temperature change information can be used to develop a prior probability distribution for the climate sensitivity, which when applied to an analysis of a paleo reconstruction, reduces the 5–95% range of climate sensitivity from the combined analysis of all proxy reconstructions used to 1.5 to 6.2°C (Hegerl et al., 2005b). Thus independent estimates, when properly combined in a Bayesian analysis, can provide a tighter constraint. Nonetheless, these results must be interpreted with caution since the exact upper limit still is still somewhat dependent on the details of the analysis, and paleoclimatic reconstructions are still being assessed for the realism of their variance (see Chapter 6).

9.6.2.2 Inferences about climate sensitivity based on the Late Maunder Minimum

Another approach to estimating the climate sensitivity is to compare different climate states. The Late Maunder Minimum (approximately 1675–1715) provides one such opportunity. During this period, sunspots were generally missing and net radiative forcing relative to present is estimated to have been close to –1.8 W m^{-2} (Section 9.2.1.3).

Different NH temperature reconstructions (Figure 6.8) yield estimates of cooling relative to the late 20th century ranging from –0.45°C (Mann et al., 1999, annual value for the Northern Hemisphere) to approximately –1.0°C (Moberg et al., 2005). Model studies (Figure 6.8) for this time period are discussed in Chapter 6, and generally yield a total cooling of –1°C to –1.5°C or greater. The radiative forcing at this period is discussed in 9.2.1.3, and is approximately –1.63 W/m^2 (relative to the present) with large uncertainties. For the approximately 50 year time period associated with the late Maunder Minimum, which was without large forcing trends, the model results may be close to radiative balance (Rind et al., 2004). However, some of the forcing during the present period is not yet realized in the system (~0.85 W m^{-2}, Hansen et al., 2005) yielding only about 0.78 W m^{-2} radiative forcing that should have been expressed in the system by the present time relative to the late Maunder Minimum. The climate sensitivity of the GCMs used for these simulations is on the order of 2.2°C (or higher, depending on the model used). Were this to have resulted in a temperature change of about –0.45°C (as in the Mann et al., 1999 reconstruction), it would imply a best-guess climate sensitivity of about 2.1°C. On the other hand, using the higher estimated cooling of 1.0°C results in an estimate of climate sensitivity up to 4.7°C (updated from Rind et al., 2004), which is broadly consistent with estimates obtained from the full millennial record (Section 9.6.2.1) and previous IPCC estimates. However, the substantial uncertainties in the total rate of forcing, past temperatures and the amount of unrealized warming makes it difficult to formally infer a probability density function for climate sensitivity from this time period.

9.6.2.3 Inferences about climate sensitivity based on the Last Glacial Maximum

The Last Glacial Maximum featured reduced greenhouse gases relative to pre-industrial levels, increased land ice as well as altered vegetation and increased atmospheric aerosols (dust primarily), while solar insolation change was negligible. Overall, the radiative forcing during the LGM is estimated to be approximately –6.1 to –10.6 W m^{-2} relative to pre-industrial (see Section 9.2.x.x).

Coupled atmosphere-ocean models run for the climate of the Last Glacial Maximum, and forced with changes in greenhouse gases and ice sheets produced global cooling estimates from 3.1°C–5.2°C when forced with changed greenhouse gases and ice sheets (Masson-Delmotte et al., 2005; see Section 9.3.3). (Note that this change may not reflect the full uncertainty in LGM temperature change, since not all forcings may be represented and uncertainties are large, see chapter 6). The radiative forcing change for the forcings and boundary conditions incorporated in the PMIP-2 LGM experiment is 4.1–7.2 W m^{-2}, resulting in a climate sensitivity estimate of 1.5 to 4.4°C for CO_2 doubling. The 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.

As with estimates of climate sensitivity based on the instrumental period, the consistency between simulations and paleo observations can be used to determine a probability density function for climate sensitivity. The sensitivity to CO_2 doubling of each model with each parameter setting is calculated by simulating the double CO_2 equilibrium climate change. Then, the probability of the different parameter settings (and with it, the different equilibrium sensitivities associated with these parameter settings) is determined from the model's ability to simulate LGM climate. Annan et al. (2005b) 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 ranges 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. The authors found a strong relationship between climate sensitivity and cooling in LGM tropical sea surface temperatures. Their results suggest that sensitivities in excess of 5.5°C are unlikely given observed estimates of LGM cooling and the relationship between tropical SST and sensitivity in their model.

The link between LGM cooling and climate sensitivity is further examined in another perturbed physics ensemble, by Schneider von Deimling et al. (2005). The authors vary 11 ocean and atmospheric parameters in 1,000 member ensemble simulations of the CLIMBER-2 EMIC (see Table 8.8.2). As with Annan et al. (2005), these authors also find a pronounced relationship between climate sensitivity and tropical SST changes in the LGM. However their relationship is different from that of Annan et al (2005b) implying a low probability for sensitivities exceeding 3.8°C, while Annan et al. (2005) find higher sensitivities consistent with data. The discrepancy between the inferred upper limits arises from structural differences between the models used, illustrating the importance of sampling uncertainties within different model frameworks to establish the robustness of simulated relationships between observables and future changes. Therefore, we conclude that these results generally support the estimates of climate sensitivity from the instrumental period and climate models, but are presently unable to constrain them further.

9.6.3 Summary of Observational Constraints for Climate Sensitivity

9.6.3.1 Caveats and uncertainties

Any constraint of climate sensitivity obtained from observations must be interpreted in light of the underlying assumptions, namely the choice of prior distribution for each of the model parameters (see Appendix 9.B) and the uncertainties that are accounted for in the approach. For example, since ocean heat uptake parameters are poorly constrained from the observations and the meaning of "effective diffusivity" is different between models (Knutti et al., 2002, 2003; Forest et al., 2002; Sokolov et al., 2003), the choice of the parameter range explored will influence the results. Neglecting sources of uncertainty in these estimates will generally result in unrealistically small uncertainties and overly narrow ranges on climate sensitivity that do not reflect the true uncertainties in our knowledge of this parameter. 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 which also affect the 5–95% range. Furthermore, structural uncertainties in the models that simulate changes with different climate sensitivities (such as problems in 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 dominant uncertainties (climate sensitivity, radiative forcing, mixing of heat into the ocean) have been taken into account in most studies (Table 9.2.1), but the following caveats must still be kept in mind. First, some processes and feedbacks might be poorly represented or missing in simple or intermediate complexity models. The use of 'effective' climate sensitivity (the equilibrium climate sensitivity a model would have assuming that feedbacks remained constant at the value found at any given point of the transient response) assumes that the climate sensitivity is truly constant in time. However some authors (e.g., Senior and Mitchell, 2000; Boer and Yu, 2003) have shown that the effective climate sensitivity varies substantially 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. A tight constraint on the sensitivity of the LGM climate is difficult to obtain because of the present uncertainty regarding tropical temperature changes.

9.6.3.2 Summary of observational constraints on climate sensitivity

Despite these uncertainties, which are accounted for to differing degrees in the various studies, estimates of equilibrium climate sensitivity show some encouraging similarities (Figure 9.6.1). This increases confidence in individual findings. Additionally, results that assess the realism of models with different sensitivity by each model's ability to simulate ice age conditions and studies using comprehensive climate models are similar, further increasing confidence. Most studies find a lower 5% limit of 1°C or greater. We conclude that it is very likely that climate sensitivity exceeds 1°C. This lower bound is consistent with the view that the sum of all atmospheric feedbacks affecting climate sensitivity is very likely positive. In most studies, the most likely value of climate sensitivity is between 1 and 4°C, averaging around on the order of 2–3°C, and rarely outside the range of 1.2 to 4°C, leading credibility to mainstream climate models used to simulate future climate change.

However, for probabilistic forecasts of future climate with constant radiative forcing, the upper tail of estimates of climate sensitivity is important. The upper 95% limit for equilibrium climate change ranges from 5°C to 10°C, or greater depending upon the approach taken, the number of uncertainties included, and specific details of the prior distribution that was used. This is, in part, caused by uncertainties and nonlinearities in forcings and response. For example, a high sensitivity cannot be ruled out because it is possible that a high aerosol forcing could nearly cancel greenhouse gas forcing. For the last millennium, uncertainties in proxy records and forcing and the nonlinear connection between sensitivity and the response to volcanism prohibit tighter constraints. For the Last Glacial Maximum, the present uncertainty about the connection between tropical SST changes and climate sensitivity in a model also makes a tight constraint difficult to obtain. Thus most studies that use a simple uniform prior on the sensitivity are not able to exclude sensitivities beyond the traditional 1.5 to 4.5°C range. However, the transient climate response, which may be more relevant for climate change in the near-term future, is easier to constrain since it relates more linearly to observables. This magnitude determines future climate change under CO2 concentrations that peak and then decline.

[START OF QUESTION 9.1]

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

Almost any weather event might occur by chance, in a climate unmodified by human activities as well as in a changed climate. However, it is possible that human influences on climate might have changed the risk of occurrence of specific extreme events (e.g., a heat wave, a widespread flooding event or a hurricane). Because extreme events are, by definition, rare it is difficult to quantify their risk and attribute the causes of changes in risk. Nevertheless, there is growing evidence of human-induced warming on continental and sub-continental scales, suggesting a possible increase in the risk of heat waves. An analysis of European mean summer temperatures indicates that human influences could have more than doubled the risk of a summer of the intensity of 2003. Quantifying the possible influence of climate change on other types of individual extreme events, including hydrological events, such as floods, is likely to prove even more difficult than is the case with temperature extremes, because of greater natural variability, lower spatial coherence and greater uncertainties in simulated climate change in rainfall.

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 (very likely the hottest in Europe for at least 500 years), the hurricanes that made landfall during the intense North Atlantic hurricane season of 2004, and the extreme daily rainfall in Mumbai, India. in July 2005. Could a human influence such as increased atmospheric greenhouse gas concentrations have “caused” any of these events?

Such a question is ill-posed. We might not be able to say that a particular weather event, given the right combination of meteorological factors, would not have happened in the absence of human-induced climate change. For example, factors that contributed to the very warm European summer of 2003 included dry soil (which leaves more solar energy available to heat the land because less energy is consumed by evaporating moisture from the soil), a persistent blocking high pressure system associated with anticyclonic subsidence that restricted vertical mixing of strongly heated surface air, and very clear skies. Such factors could arise in a climate unmodified by human influences.

Our question can, however, be re-posed as “How has anthropogenic forcing changed the probability of occurrence of extreme weather events such as heat waves?” This question can be addressed, for example for the 2003 European heat wave, by studying the characteristics of European summers in a climate model, either forced solely with natural factors such as volcanic activity and changes in solar radiation, or by both anthropogenic and natural factors. Experiments of this type indicate that it is very likely that anthropogenic forcing is responsible for a substantial fraction of the summer warming trend observed over Europe in recent decades, and thus an increase in the likelihood of extremely warm summers. These experiments indicate that 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 impacts, such as a successive number of very warm nights in an urban area such as Paris, the type of meteorological indicator most closely related to increased mortality in the summer of 2003.

The value of a risk-based approach (“is there an increase in the probability of an event that is attributable to 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. Some care must be taken in assessing changes in risks of extreme events. Careful statistical analyses are required, since the risk of individual extremes, such as very hot days exceeding some threshold, could change due to changes in variability as well as changes in the mean climate. Such analyses rely on model-based estimates of variability, and thus an important additional requirement is that climate models adequately represent climate variability.

The same risk based approach could be adopted to examine possible changes in the frequency of extreme hydrological events such as heavy rainfalls or floods. However rainfall is much more variable from one location to the next than temperature, increasing the difficulty of ascribing causes. High resolution modelling is likely to be required to capture details of the climate change signal which may be difficult to identify above a high level of natural variability noise. Also, rainfall is less reliably simulated than temperature. As the signal of climate change becomes stronger in future, 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. There is already some evidence of increases in extreme rainfall events in at least some regions. However there is no convincing evidence as yet that there is a clear link between these increases and increasing greenhouse gas concentrations in the atmosphere.

[END OF QUESTION 9.1]

[START QUESTION 9.2]

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

Unlikely. There has been a rapid warming over the last century that is very unusual relative to other periods in the last millennium according to estimates deduced from proxy records of temperature such as tree rings and ice cores. This rapid warming is consistent with our physical understanding of how the climate would respond to a rapid increase in greenhouse gas concentrations. Observed patterns of recent temperature change 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. Climate models provide good simulations of the last 100 years when they include the dominant external forcings that were experienced by the climate system during this period. Moreover, models fail to reproduce the warming observed in recent decades when anthropogenic forcings are not included. Climate models almost never simulate warming rates of the magnitude observed over the 20th century when external forcings of the climate system are held constant, even in very long multi-millennium simulations.

The warming seen over the 20th century is consistent with scientific understanding, based on simple energy balance considerations, of how the climate is expected to respond to perturbations from forcings of the climate system, both man-made and natural. 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 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. Explosive volcanic eruptions periodically ejected large amounts of dust and sulphate aerosol high up in the atmosphere temporarily shielding the earth and reflecting solar radiation back to space. Solar output varied on an 11 year cycle and could also have had longer term variations.

Although internal variations in the climate system can cause changes in global temperature from decade to decade, much of the multi-decadal changes in global mean temperature over the 20th century are likely to have been caused by external forcings. Short-lived decreases in global temperature followed volcanic eruptions such as Pinatubo in 1991. The levelling off of global mean temperatures in the 1950s and 1960s accompanied increases in sulphate aerosols, which have the potential to cool the planet by reflecting radiation directly to space and by making clouds more reflective and longer lasting. There was global mean warming in the early part of the 20th century, a period during which greenhouse gas concentrations had started to rise, volcanic activity was reduced, and when solar output was likely increasing. The warming observed since the 1970s has occurred in a period when greenhouse gas forcing has dominated over all other forcings.

The patterns of warming 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 anthropogenic forcing but also differ in many respects from patterns of temperature change associated with the most important modes of internal variability, such as the El Niño Southern Oscillation.

Numerous experiments have been conducted with coupled ocean-atmosphere climate models, forced with natural and/or anthropogenic factors, to determine the likely causes of the 20th century climate change. These experiments, conducted with a variety of models, indicate that the warming of global surface temperature cannot be reproduced by the models when they are forced only with natural variations in external forcing (i.e., variations in solar radiation and volcanic aerosols). However, models are able to simulate the observed warming when they include all of the most important forcings, including forcings from anthropogenic sources (changes in greenhouse gases, stratospheric ozone, and industrial aerosols) as well as natural forcings. Long multi-century climate simulations have also been used to estimate the variability of the climate system in the absence of natural and anthropogenic external forcing changes. Century scale temperature changes that result from internal climate variability, as simulated by models, are quite small relative to the strong 20th century warming, suggesting that internal climate system mechanisms are very unlikely to have produced the observed warming.

An important source of uncertainty arises from the incomplete knowledge about the temporal and spatial variations in forcings, such as those associated with anthropogenic aerosols, used in model experiments. In addition, the climate models themselves are also imperfect. However models exhibit a common pattern of response to anthropogenic greenhouse forcing including 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 forcings, for example from increasing solar output which has a different vertical pattern of response. In addition, the different time histories of anthropogenic and natural forcings help to distinguish between the responses. Such considerations increase confidence that warming observed over the last 50 years was dominated by anthropogenic rather than natural factors.

Reconstructions of Northern Hemispheric temperature of the last 1000 years that are 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 a second line of evidence that natural variability cannot explain the 20th century warming. While recent work has shown that there is some uncertainty in the level of past temperature variability, all reconstructions, including the most variable, strongly suggest that the warming observed during the 20th century is stronger than any other century-scale global temperature change estimated to have occurred over 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. Temperature variations during the last 150 years of these records can also be explained if anthropogenic forcing is also taken into account. Thus, despite some uncertainty in these records, they provide a picture of the role of external influences on the climate system and the unusual nature of the 20th century that is consistent with the results of the analysis of the 20th century instrumental record.

[END OF QUESTION 9.2]

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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 $\mathbf{y} = \mathbf{X}\mathbf{a} + \mathbf{u}$ where vector **y** is a filtered version of the observed record, matrix **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 $\mathbf{u}$ is a realization of internal climate variability. Vector $\mathbf{u}$ is 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 **X** typically contains signals that are estimated with either a CGCM, an atmospheric general circulation model (AGCM; see Sexton et al., 2001, 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 long control simulations with CGCMs (i.e., without external forcing) 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, 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 control simulation.

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 deductive reasoning (to determine whether there is evidence that 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; Berliner et al., 2000; Min et al., 2004; Min and Hense, 2005a; Schnur and Hasselmann, 2005; Lee et al., 2005a; see also Kettleborough, 2005).

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). 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. Also, Bayesian inferences are probabilistic (i.e., based on the posterior likelihoods of detection and attribution), which means that they can better feed into decision making processes that balance risks and benefits.

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 (2005a). In contrast, Berliner et al. (2000) and Lee et al. (2005a) 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 different 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. That approach is explained in Section 9.6. 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 OAGCMs. 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. (2005b). 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 distance of any residual to 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; 2004; Hegerl et al., 2005b) 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 function $p(T_{obs} \mid \theta)$, known as the likelihood function, which 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 $p(\theta \mid T_{obs})$ on the parameters. That is, one calculates $p(\theta \mid T_{obs}) \propto p(T_{obs} \mid \theta) \cdot p(\theta)$ where the product on the right is normalized so that the integral of $p(\theta \mid T_{obs})$ 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 $p(\theta \mid T_{obs})$ is integrated over the range of free parameters other than climate sensitivity to yield a marginal posterior probability density function for climate sensitivity. Similar integrations must be performed to obtain marginal probability density functions on the other parameters.

The prior distribution $p(\theta)$ that is used in this calculation is chosen to reflect prior knowledge (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, choice of prior bounds can be somewhat subjective. In the case of climate sensitivity, some studies (e.g., Forest et al., 2002, 2004), have used 0°C and 10°C as lower and upper bounds respectively.

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 assessment presented in Section 9.6 concludes that inferences on climate sensitivity, for example, remain sensitive to the choice of prior, indicating that observations do not yet contain sufficient information to constrain this parameter. The approach used in most studies is to place uniform priors on the parameters of interest, such as the climate sensitivity. However, observable properties of the climate system appear to scale with the strength of atmospheric feedbacks rather than 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 climate sensitivity, and therefore results in narrower posterior likelihood ranges on the climate sensitivity that exclude very high sensitivities. The decision as to how to sample a given climate parameter (e.g., whether to place a prior on the climate feedback as opposed to the climate sensitivity) ultimately depends upon the application (e.g., whether the objective is to constrain projections of future climate change, or the equilibrium climate sensitivity).

Alternatively, expert opinion can also be used to construct priors, as done by Forest et al. (2002; 2005). Note, however, that expert opinion may be overconfident (Risbey and Kandlikar, 2002) and if this is the case, the posterior distribution may again 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. Given the limited data available, it is impossible to separate prior beliefs about climate system properties from knowledge of relevant climate observations.

Tables

Table 9.1.1. A synthesis of climate change detection results. **a)** Surface and atmospheric temperature evidence. Note that while individual global and regional detection studies identify an anthropogenic signal in observations relative to estimates of internal variability, our likelihood assessments are necessarily conservative because the number of studies assessing the impact of forcing and model uncertainty remains limited. **b)** Evidence from other variables.

a)

Result	Region	Likelihood	Main Sources of Uncertainty and Certainty
Surface temperature			
Warming during the past half century is not solely due to known sources of internal climate variability	Global	Virtually Certain (>99%)	Anthropogenic change detected with very high significance levels in surface temperature (much less than 1%) and the upper ocean. Observed change is very large relative to climate model simulated internal variability. Variability simulated by models consistent with variability estimated from instrumental and paleo records. (Sections 9.4.1.2, 9.4.1.4, 9.5.1.2, 9.3.4.2),
Warming during the past half century is not solely due to known natural causes	Global	Very Likely (90–99%)	This warming took place at a time when non-anthropogenic factors would likely have produced cooling. The combined effect of known sources of anthropogenic forcing would have been very likely to produce a warming. Main uncertainty from forcing and internal variability estimates (Sections 9.4.1.2, 9.4.1.4)
Early 20th century warming due in part to anthropogenic and natural external forcing.	Global	Likely (66–90%)	Natural forcing and response uncertain, some observational uncertainty in early 20th century trend. (Sections 9.3.4.1, 9.4.1.4).
Greenhouse gas warming greater than observed during latter half of 20th century.	Global	Likely (66–90%), bordering on Very Likely (90–99%)	Separation of response to non-greenhouse gas (particularly aerosol) forcing from greenhouse gas forcing varies between models (although greenhouse gas response estimates consistent between models). (Section 9.4.1.4)
Greenhouse warming detected on all continents except Antarctica (latter half 20th century).	Continental to sub-continental	Likely (6690%)	Sampling (lower signal to noise ratio on subcontinental scale – multi-signal results less consistent as a result). (Section 9.4.2)
Late 20th century warmer than at any other time in past 1000 years.	NH (mostly extra-tropics)	Likely (66–90%)	Uncertainty in reconstructions, particularly in the magnitude of past variations. (Chapter 6)
Preindustrial temperatures significantly affected by natural external forcing (~1000–1850)	NH (mostly extra-tropics)	Very Likely (90–99%)	Uncertainty in reconstructions and past forcings; however, 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, Australia	Likely (66–90%).	Model uncertainties and limited observational coverage. 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. Anthropogenic influence in temperature of warmest day was not detected. (Section 9.4.3)

Free atmosphere changes			
Tropopause height changes detectable and attributable to anthropogenic forcing (latter half 20th century)	Global	Likely (6690%)	Observations rely on reanalysis data, although mostly during the better constrained satellite period. Modelled changes result mainly from tropospheric waming from greenhouse gas increases and ozone-induced stratospheric cooling, although uncertainties remain concerning influence of stratospheric ozone changes on tropopause height (Section 9.4.4.2).
Tropospheric warming detectable and attributable to anthropogenic forcing (latter half 20th century)	Global	Likely (66–90%)	Model and observational. Some models may respond too strongly to anthropogenic forcing, although simulated response of other models now consistent with the recently reconciled satellite records. The radiosonde record continues to be uncertain. (Section 9.4.4)

b)

Result	**Region**	**Likelihood**	**Main Sources of Uncertainty and Certainty**
Ocean changes			
Detectable anthropogenic warming of upper ocean (latter half 20th century)	Global (but with limited sampling in some regions)	Very Likely (90–99%)	Observations and models. Limited sampling of ocean interior makes it difficult to assess model performance. Greater confidence in results for well-sampled, upper 300m layer of ocean. (Section 9.5.1.2)
Circulation			
Sea level pressure shows detectable anthropogenic signature (latter half 20th century)	Global	Likely (66–90%)	Models and observations. Models underestimate the observed NH changes for reasons that are not understood, and few studies. Simulated responses in the SH to ozone concentration changes are consistent with observations (Section 9.5.2.4).
Increase in tropical cyclone intensity and duration since the 1970s due to anthropogenic forcing	Tropical regions	Inconclusive at this stage.	Models (insufficient resolution) and sampling (signal to noise ratio likely low). Process studies anticipate only modest changes in intensity by end of 21st century. Observational evidence indicates that increases cyclone intensity and duration since the 1970s are consistent with SST and atmospheric water vapour increases. (Box 3.4 and Section 9.5.2.6)
Precipitation, Drought, Runoff			
Rainfall changes show detectable volcanic signal (latter half 20th century)	Global land areas	Likely (66–90%), bordering on as Likely as Not (33–66%)	Models, forcings and observations. Model and observed response different in amplitude. Model response not detectable in observations for all models. However, result supported by theoretical understanding. (Section 9.5.3.2)
Increases in heavy rainfall consistent with anthropogenic forcing (latter half 20th century)	Global land areas (limited sampling)	About as Likely as Not (33–66%)	Results vary between regions and studies. Changes in some regions consistent with expectations. 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 (latter half 20th century)	Global land areas	About as Likely as Not (33–66%), bordering on Likely (66–90%)	Not formally detected..Observed trends inconsistent with simulated natural internal variability and consistent with simulated response to external forcing in one modelling study. Studies suggest that regional droughts are linked either to SST changes that, in some instances, may be linked to anthropogenic aerosol forcing (e.g., Sahel) or a circulation response to anthropogenic forcing (e.g., SW Australia). Models, observations and forcing all contribute uncertainty. (Section 9.5.3.2)
Increase in continental runoff due to anthropogenic forcing (latter half 20th century)	Global land areas	About as Likely as Not (33–66%), probably >50%	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).
Cryosphere			
Reductions in NH sea-ice extent due to anthropogenic forcing (latter half of 20th century)	Arctic	Likely (66–90%)	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 simulations with another climate model. (Section 9.5.4.1)
Glacier retreat due to anthropogenic forcing (20th century)	Global	Very Likely (90–99%)	Formally detected (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. (Section 9.5.4.3)

Table 9.2.1. Results from key studies on observational estimates of the equilibrium climate sensitivity parameter α (in °C/CO_2 doubling) and the total or net fossil fuel-related aerosol forcing "aer" (in W m^{-2}). All assume non-negative sensitivity to avoid unstable climate. The following uncertainties are incorporated in some studies: κ refers to ocean diffusivity, ε_{aer} to aerosol uncertainty, ε_{nat} to natural forcing uncertainty, ε_{obs} observational uncertainty. "Noise" refers to internal climate variability. Note that the aerosol range refers to the net fossil-fuel related aerosol range, which tends to be all forcings not accounted for by greenhouse gases, ozone or natural forcing that project onto the pattern associated with fossil fuel aerosols, and it includes all unknown or not explicitly considered forcings. "Mode" refers to the most likely value (peak of the probability PDF). Values are taken from publications and depend on ranges considered, priors etc.
Key to forcings: G: greenhouse gases; Sul: direct sulphate aerosol effect; Suli: indirect sulphate; OzT: tropospheric ozone; OzS: stratospheric ozone; Vol: volcanism; Sol: solar, BC+OM: black carbon and organic matter

Study	Data	Model	Forcing	Free Parameters	α Range 5–95% [°C]	Mode	Aer 5–95% Range [W/m^2]	Incorporated Uncertainties
Forest et al., 2002	Upper air, surface and deep ocean space-time 20th C Temp. (diag 2nd half)	2-D OAGCM	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	–0.3 to –0.95* –0.25 to –0.9 with expert prior	ε_{obs} , noise (OAGCM)
Forest et al., 2005	"	"	G, Sul, Sol, Vol, OzS, land surface changes	"	2.4–9.2 2.2–5.2 with expert prior	3.0	–0.16 to –0.7 –0.05 to –0.62 with expert prior	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, Suli	α , choice of radiative forcing factors	1.0 to 9.3 p>54% % that outside 1.5–4.5	1.2	–0.54 to –1.3	Noise (bootstrap residual)
Knutti et al., 2002, 2003	Global mean ocean heat uptake 1955–1995, mean SAT inc. 1900–2000	EMIC, (+ 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 outside 1.5–4.5	Median 4.9	0 to –1.2 ind. aerosol, –0.6 to –1.7 total aer	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 et al., 2001), Sol, Vol	Directly estimated	1.1 to inf with BC, 1.6 to inf. Without	2.1	Derived from Stott et al., 2000: –0.4 to –1.6	ε_{obs} , ε_{forc}
Frame et al., 2005	Global change in surface temperature	EBM	G, Sul, Nat?	Range of κ consistent with ocean warming	1.2 to 11.8 (sampled to 20) (0.4 to 4.0 if unif. feebacks)	2.0	Aerosol response separated out	Noise, forcing uncertainties

Forster and Gregory, 2005	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	Changes not important in period chosen	ε_{obs} , forcing uncertainty discussed
Hegerl et al., 2005b	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	Not constrained to 1850, not well to 1960	ε_{obs} , ε_{nat} , ε_{aer} , noise (from residual)

Notes:
[*]Implicitly include OZT and fossil fuel BC+OM since these were excluded from the study.

Table 9.3.1. Detection and attribution results for a range of paleoclimatic records of the last millennium. “Y” or “N” indicates that the response to external forcing is, or is not, detectable; “small” or “large” indicates that signals are significantly under- or overestimated. The bottom row gives the standard deviation of the (decadally smoothed for annual data) residual and, in parentheses, the amount of decadal variance explained by external forcing. From Hegerl et al. (2003).

Record Period	Briffa et al., (2001) 1400–1940	Crowley -Lowery, 2000 1400–1960	Mann et al., 1999 14001960	Esper et al., 2002 1400–1960	Crowley-Lowery, 2000 1000–1960
Volcanic	Y	Y	Y small	Y	Y
Solar	N, small	N, small	N (Y for periods)	N, small	N (Y from 1100 on)
Ghg + aer	Y	Y	N small (Y to 1980)	Y, large	Y
Residual std	0.09 (57%)	0.08 (77%)	0.07 (49%)	0.13 (67%)	0.10 (57%)