Working Group I Contribution to the
Intergovernmental Panel on Climate Change
Fourth Assessment Report

Climate Change 2007: The Physical Science Basis

Technical Summary

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TS.1 INTRODUCTION

In the last 6 years since the IPCC's Third Assessment Report (TAR) on climate change, significant progress has been made in understanding past, recent, and future climate change. These advances have arisen from: large amounts of new data, more sophisticated analyses of data, improvements in the simulation of physical processes in climate models, and more extensive exploration of uncertainty ranges in model results. The increased confidence in climate science provided by these developments, as well as advances in probabilistic approaches, are over-arching themes in the Working Group I contribution to the IPCC's Fourth Assessment Report.

While the progress in climate science summarized in this report provides the present status of new and important policy-relevant information, the complexity of the climate system and the multiple interactions that determine its behaviour impose limitations on our ability to predict fully the future course of Earth's climate. There is still an incomplete physical understanding of many components of the climate system and their role in climate change. These include aspects of the roles played by clouds, the cryosphere, the oceans, land-use, and couplings between climate and biogeochemical cycles. The areas of science covered in this report continue to undergo rapid progress and it should be recognized that the present assessment reflects scientific understanding based on the peer-reviewed literature available in early 2006, as required by IPCC procedures.

The key findings of the IPCC Working Group I assessment are presented in the Summary for Policymakers. This Technical Summary provides a more detailed overview of the scientific basis for those findings and provides a road map to the chapters of the underlying report. The structure of the Technical Summary is as follows:

- Section 2: an overview of current scientific understanding of the natural and anthropogenic drivers of changes in climate;
- Section 3: an overview of observed changes in the climate system (including the atmosphere, oceans and cryosphere) and their relationships to physical processes;
- Section 4: an overview of explanations of observed climate changes based on climate models and the extent to which climate change can be attributed to specific causes;
- Section 5: an overview of projections for both near- and far-term climate changes including the time scales of responses to changes in forcing, and probabilistic information on future climate change;
- Section 6: a summary of the most robust findings and the remaining key uncertainties in current physical climate change science.

Each paragraph in the Technical Summary reporting substantive results is followed by a reference in square brackets to the corresponding chapter section(s) of the underlying report where the detailed assessment of the scientific literature and additional information can be found.

BOX TS.1.1: TREATMENT OF UNCERTAINTIES IN THE WORKING GROUP I ASSESSMENT

The importance of consistent and transparent treatment of uncertainties is clearly recognized by the IPCC in preparing its assessments of climate change. The increasing attention given to formal treatments of uncertainty in previous assessments is addressed in Chapter 1, Section 1.6. To promote consistency in the general treatment of uncertainty across all three Working Groups, authors of the Fourth Assessment Report have been asked to follow a brief set of guidance notes on determining and describing uncertainties in the context of an assessment[1]. This box summarises the way in which those guidelines have been applied by Working Group I and covers some aspects of the treatment of uncertainty specific to material assessed here.

Uncertainties can be classified in several different ways according to their origin but at least two different types should be recognized. *Value uncertainties* arise from the incomplete determination of particular values or results, e.g. when data are inaccurate or not fully representative of the phenomenon of interest. *Structural uncertainties* arise from an incomplete understanding of the processes that control particular values or

[1] See http://www.ipcc.ch/activity/uncertaintyguidancenote.pdf

results, e.g. when the conceptual framework or model used for analysis does not include all the relevant processes or relationships. Value uncertainties are generally estimated using statistical techniques and expressed probabilistically. Structural uncertainties are generally described by giving the authors' collective judgment of their confidence in the correctness of a result. In both cases estimating uncertainties is intrinsically about describing the limits to knowledge and for this reason involves expert judgment about the state of that knowledge.

Although the terminology used is not always consistent, the science literature assessed here uses a variety of generic ways of categorizing uncertainties. Random errors have the characteristic of decreasing as additional measurements are accumulated, whereas systematic errors do not. In dealing with climate records considerable attention has been given to the identification of systematic errors or unintended biases arising from data sampling issues and methods of analysing and combining data. Specialized treatments of uncertainties have been developed for the detection and attribution of climate change and for producing probabilistic projections of future climate parameters. These are summarised in the relevant chapters.

The uncertainty guidance provided for the Fourth Assessment Report draws, for the first time, a careful distinction between levels of confidence in our scientific understanding and likelihoods of specific results. This allows authors to express high confidence that an event is very unlikely (e.g. rolling six with a dice three times in a row), or about as likely as not (e.g. a tossed coin coming up heads). Confidence and likelihood as used here are distinct concepts but are often linked.

The standard terms used to define levels of confidence in this report are as given in the Uncertainty Guidance Note, viz:

Terminology	Degree of confidence in being correct
Very High confidence	At least 9 out of 10 chance of being correct
High confidence	About 8 out of 10 chance
Medium confidence	About 5 out of 10 chance
Low confidence	About 2 out of 10 chance
Very low confidence	Less than 1 out of 10 chance

Chapter 2 of this report uses a related term "level of scientific understanding" when describing uncertainties in different contributions to radiative forcing. This terminology is used for consistency with the Third Assessment Report and the basis on which the authors have determined particular levels of scientific understanding uses a combination of approaches consistent with the uncertainty guidance note as explained in detail in Chapter 2, Section 2.9.2 and Table 2.11.

The standard terms used in this report to define the likelihood of an outcome or result where this can be estimated probabilistically are:

Terminology	Likelihood of the occurrence/ outcome
Virtually certain	> 99% probability of occurrence
Very likely	> 90% probability
Likely	> 66% probability
About as likely as not	33 to 66% probability
Unlikely	< 33% probability
Very unlikely	< 10% probability
Exceptionally unlikely	< 1% probability

In order to provide a more specific assessment of detection and attribution of key aspects of climate change Chapter 9 of this report augments the likelihood scale above with additional terms "*Highly likely*" to indicate a greater than 95% likelihood of occurrence, and "*More likely than not*" to indicate a greater than 50% likelihood.

Where values are specified in this report as a central estimate with a plus/minus range, then by default the range represents a 95% (2-σ) confidence interval. Exceptions to this are noted in the text.

TS.2 CHANGES IN HUMAN AND NATURAL DRIVERS OF CLIMATE

The Earth's global mean climate is determined both by incoming energy from the Sun and the properties of the Earth and its atmosphere, particularly the reflection, absorption, and re-emission of energy within the atmosphere and at the surface. Thus while changes in received solar energy (e.g., caused by variations in the Earth's orbit around the Sun) inevitably affect the Earth's energy budget, the properties of the atmosphere and surface, which themselves may be affected by climate feedbacks, are also important. The important roles of feedbacks are evident in the paleoclimatic record of changes in the ice sheets and their association with pre-historic climate changes.

Changes have been documented in several agents that alter the global energy budget of the Earth and therefore can cause the climate to change. Among these are increases in greenhouse gas concentrations that act primarily to increase atmospheric absorption of outgoing radiation, and increases in aerosols (microscopic airborne particles or droplets) that act to reflect and absorb incoming solar radiation. Changes in such forcing agents cause radiative forcing of the climate system[2]. The agents can differ considerably from one another in terms of the magnitudes of forcing, as well as spatial and seasonal features. Positive and negative radiative forcings respectively contribute to increases and decreases in globally averaged temperature. This section updates the understanding of the estimated anthropogenic and natural radiative forcings. New insights into the effects of forcing agents such as aerosols on precipitation are also highlighted.

The response of global climate to radiative forcing is complex due to a large number of feedbacks within the climate system. The response involves both positive and negative feedbacks, with the result that each process has to be modelled accurately in order to estimate the overall response. While water vapour is a strong greenhouse gas, it is well established that many of its changes are due to climate feedback mechanisms, and are therefore part of the climate response rather than a radiative forcing. Changes in clouds linked to climate changes are another important climate feedback. Important progress has been achieved in understanding the relationships between global mean radiative forcing, feedbacks, and climate response, and this is briefly summarized in this section. A more detailed analysis of model evaluation is presented in Sections 4 and 5.

TS.2.1 GREENHOUSE GASES

The dominant factor in the radiative forcing of climate in the industrial era is the increasing concentration of various greenhouse gases in the atmosphere. Several of the major greenhouse gases occur naturally but their increasing atmospheric concentrations can be attributed to human activities. Other greenhouse gases are solely the result of human activities. The contribution of each greenhouse gas to radiative forcing over a particular period of time is determined by the change in its concentration in the atmosphere over that period and the effectiveness of the gas (per ppb) in perturbing the radiative balance. The atmospheric concentrations of the different greenhouse gases considered in this report vary by more than 8 orders of magnitude (factor of 10^8), and their radiative efficiencies vary by more than 4 orders of magnitude (factor of 10^4); this reflects the enormous diversity in their properties and origins.

The concentration of a greenhouse gas in the atmosphere or, more important, the change in that concentration since pre-industrial times (defined as the period since 1750), is a function of the history of past emissions and the removal rate or atmospheric lifetime of the gas.

Long-lived greenhouse gases (LLGHGs) are chemically stable and persist in the atmosphere over time scales of a decade to centuries or longer, so that their emission has a long-term influence on climate (e.g., carbon dioxide, methane, nitrous oxide). Because these gases are long-lived in the atmosphere, their global concentrations can be accurately estimated from data at a few locations. Carbon dioxide (CO_2) is cycled between the atmosphere, oceans and land biosphere and its removal from the atmosphere involves a range of processes with several different timescales. However, a portion of the anthropogenic enhancement to CO_2 concentrations is expected to persist in the atmosphere for thousands of years.

[2] *Radiative forcing* is a measure of the influence a factor has in altering the balance of incoming and outgoing energy in the Earth-atmosphere system and is an index of the importance of the factor as a potential climate change mechanism. It is expressed Watts per square meter (W m^{-2}). See Glossary for further details.

Short-lived gases are generally chemically reactive and removed by natural oxidation processes in the atmosphere or by washout in precipitation (e.g., sulphate, carbon monoxide). Ozone is also a significant greenhouse gas that is formed through chemical reactions in the atmosphere. In the troposphere the human influence on ozone occurs primarily through changes in precursor gases that lead to its formation, while in the stratosphere, the human influence has been primarily through changes in ozone removal rates due to chlorofluorocarbons (CFCs) and other ozone-depleting substances.

TS.2.1.1 Radiative forcing due to carbon dioxide, methane, and nitrous oxide

Analyses of modern air samples together with air samples extracted from polar ice cores now provide detailed time series data over the past 650,000 years for CO_2, methane, and nitrous oxide (see Figure TS-1). [2.3, 6.3, 6.4, 6.5]

[INSERT FIGURE TS-1 HERE]

The total radiative forcing of the Earth's climate due to increases in the concentrations of the LLGHGs CO_2, methane (CH_4), and nitrous oxide (N_2O), and the rate of observed increase in these gases over the past century, are unprecedented in at least the last 20,000 years as shown in Figure TS-1. It is very likely that the average rate of increase in radiative forcing from these well-mixed greenhouse gases over the past decade is at least six times faster at present than at any time during the two millennia before the Industrial Era, the period for which decadally-resolved ice core data exist. The radiative forcing due to these gases has increased by about 0.25 W m^{-2} per decade in the past 50 years. The rate of increase of radiative forcing since 1950 is much larger than the rate of increase in any similar period prior to 1950. This component of radiative forcing of the global mean Earth’s climate system has the highest level of confidence of any forcing agent. [2.3, 6.4]

Increases in atmospheric CO_2 since pre-industrial times are responsible for a radiative forcing of 1.63 ± 0.16 W m^{-2}; a contribution that dominates that of all other radiative forcing agents considered in this report. The increase of CO_2 in the atmosphere has yielded the largest radiative forcing of any known forcing agent since preindustrial times. Atmospheric CO_2 concentrations increased by only 20 ppm over the 8,000 years prior to industrialization, and multi-decadal-to-centennial scale variations were less than 10 ppm. However, since 1750, its concentration has risen by over 100 ppm. Between 1999 and 2004, CO_2 increased at an average rate of more than 1.8 ppm yr^{-1}. The estimated increase in its radiative forcing since the TAR is larger than the corresponding change due to any other agent. (see Figure TS-2). [2.3, 6.4, 6.5]

[INSERT FIGURE TS-2 HERE]

Increases in atmospheric methane concentrations since preindustrial times make it an important greenhouse gas. Its current radiative forcing is second only to that of CO_2, at 0.48 ± 0.05 W m^{-2}. Atmospheric methane concentrations varied slowly between 750 and 550 ppb over the last 11,500 years, but increased by about 1000 ppb in just the last two centuries, representing the fastest changes in at least the last 80,000 years. However, since the TAR, the growth rate of atmospheric methane concentrations has decreased, and averaged 0.8 ppb yr^{-1} (0.04% yr^{-1}) for the 5-year period from 1999 to 2004 from highs of greater than 1% yr^{-1} in the late 1970s and early 1980s. [2.3, 6.4, 6.5]

Nitrous oxide continues to rise approximately linearly by about 0.8 ppb yr^{-1}, and now contributes a radiative forcing of 0.16 ± 0.02 W m^{-2}. Ice core data show that the atmospheric concentration of N_2O was almost stable for the last 11,500 years, before the onset of the industrial period. [2.3, 6.4, 6.5]

TS.2.1.2 Sources and sinks of carbon dioxide, methane, and nitrous oxide

Polar ice core data now demonstrate that current concentrations of atmospheric CO_2, CH_4 and N_2O are all substantially above the concentrations found in the longest (up to 650,000 years) records, supporting the understanding that the post-industrial rise in these gases does not stem from natural mechanisms. [2.3, 6.4]

The concentration of atmospheric CO_2 increases with the magnitude of fossil fuel and other emissions such as those from land-use change, but also depends on the rate of carbon uptake by the ocean and by the land biosphere. Over the period from 1999 to 2005, fossil fuel emissions rose from 6.5 to 7.2 GtC yr^{-1}. The temporal evolution of emissions associated with land use change is less well known but they are estimated to have contributed 5 to 38% of CO_2 growth in the 1990s. Table TS-1 shows the estimated budgets of CO_2 in recent decades. [2.3, 7.3]

Table TS-1. Global carbon budget. By convention, CO_2 fluxes leaving the atmospheric reservoir (i.e. "CO_2 sinks") have a negative sign. Numbers in parentheses are ranges. Units: GtC yr^{-1}. NA: No information available to separate.

	1980s	1990s	2000–2005
Atmospheric increase	3.3 ± 0.1	3.2 ± 0.1	4.1 ± 0.1
Emissions (fossil fuel+cement)	5.4 ± 0.3	6.4 ± 0.3	7.0 ± 0.3
Ocean-atmosphere flux	–1.8 ± 0.8	–2.2 ± 0.4	–2.2 ± 0.4
Land-atmosphere flux	–0.3 ± 0.9	–1.0 ± 0.5	–0.7 ± 0.5
Partitioned as follows			
Land use change flux	1.3 (0.3 to 2.8)	1.6 (0.5 to 2.8)	NA[3]
Residual land sink	–1.6 (–4.0 to 0.3)	–2.6 (–4.3 to –1.0)	NA

Measurements of carbon isotopes and of oxygen provide evidence showing that human activities are responsible for observed CO_2 increases. CO_2 uptake rates to the biosphere and oceans are dependent on the atmospheric CO_2 concentration and on climate. On average, 30–50% of the net removal of CO_2 from the atmosphere, and a larger fraction of the inter-annual variation, may be attributed to uptake by the terrestrial biosphere. The balance (50–70%) is taken up by the oceans. This partitioning of net CO_2 uptake by the biosphere and oceans has been determined from a variety of techniques, including use of high precision measurements of atmospheric oxygen and of the carbon isotopes in CO_2. Fluxes of O_2 and CO_2 are inversely related in plant photosynthesis, respiration and combustion of fossil fuel, whereas CO_2 uptake by the ocean is not linked to a corresponding O_2 flux. Thus changes in atmospheric O_2 and CO_2 can be used to determine the net carbon uptake (sinks minus sources) by the ocean and the land biosphere. Carbon isotopes can also be used to determine the carbon budget as different carbon fluxes have different isotopic ratios. [5.4, 7.3]

Carbon uptake and storage in the terrestrial biosphere is the net difference between uptake due to growth, reforestation and sequestration and emissions due to heterotrophic respiration, harvest, deforestation, fire, damage by pollution, and other disturbance factors affecting biomass and soils. Forests, soils and peatlands contain the largest stocks of organic matter globally, and future release of carbon from these reservoirs has the potential to significantly increase atmospheric CO_2 concentrations. During the past 25 years, large-scale clearing of forests has occurred in the tropics whereas forest areas generally increased at middle and high-latitudes during the 20th century. Re-growing forests on abandoned former agricultural lands take up significant amounts of carbon every year until they reach a statistical equilibrium. Commercially managed forests also take up carbon which, however, is balanced by losses due to harvest, hence in general do not constitute a net sink. The current carbon balance of high-latitude forests is not clear. Significant re-growth of forests appears to be taking place across Scandinavia and Russia, but it is estimated that forest fires have also increased in North America and Eurasia, perhaps in association with climate variations and trends. [7.3]

The effects of increasing CO_2 on terrestrial carbon uptake on a large scale cannot be quantified reliably at present. Plant growth can be stimulated by increased atmospheric CO_2 concentrations and nutrient deposition (fertilization effects). However, most experiments and studies show that such responses appear to be relatively short lived and strongly coupled to other effects such as availability of water and nutrients. Likewise, experiments and studies of the effects of climate (temperature and moisture) on heterotrophic respiration of litter and soils are equivocal. [7.3]

[3] Not available

Short-term (interannual) variations in CO_2 growth rate are primarily controlled by changes in the exchange of CO_2 between the atmosphere and the terrestrial biosphere, with the ocean contributing a smaller but significant fraction of the observed variability. The terrestrial biosphere flux variability is driven by climatic fluctuations affecting the uptake of CO_2 by plant growth and the return of CO_2 by the decay of organic material through heterotrophic respiration and fires. In addition, interannual variations in CO_2 emissions associated with land-use and biomass burning can be significant. Decadal and longer-term trends in the atmospheric CO_2 growth rate reflect the balance between trends in CO_2 emission rates from fossil fuel burning and increases in the total rate of removal. During the 1990s, estimates of the annual ocean-atmosphere flux of CO_2 ranged between –1.8 and –2.8 Gt yr^{-1}. El Niño Southern Oscillation (ENSO) events represent a major source of interannual variability, not only because of large changes in the net flux to the ocean linked to changes in SSTs, but also because of the high correlation between ENSO and the release of CO_2 to the atmosphere from fires. (Figure TS-3). [7.3]

[INSERT FIGURES TS-3]

Since the early 1990s the average methane growth rate has decreased. The TAR considered several reasons for observed decreases in the growth rate of atmospheric methane including a decrease in its sources, an increase in its principal atmospheric sink (the hydroxyl radical (OH)) and a combination of changes in both sources and sinks. Since the TAR experimental work published by several laboratories using two different tracers (methyl chloroform and ^{14}CO) suggests no long-term change in the global abundance of OH over 25 and 13 year records respectively. Reasons for the decline in the growth rate of methane are not well understood. [2.3, 7.4]

The current methane abundance of about 1.777 ppm is more than double its preindustrial value. Although methane growth rates have declined, observations and modelling studies show that current methane levels are due to continuing anthropogenic emissions of methane which are greater in magnitude than its natural sources. Individual sources of methane are not as well quantified as total emissions but are mostly biogenic and include wetlands, rice agriculture, biomass burning and ruminant animals with smaller contributions from industrial sources including fossil fuel-related emissions. Emissions of methane from energy use and from northern wetlands appear to have decreased, while emissions from lower latitudes have probably been increasing. Recent observations underscore a decoupling of methane concentration changes from human population growth and suggest uncertainties of the impact of global change on natural sources. [2.3, 7.4]

A feature of the slowdown in the growth rate of methane in the atmosphere over the last 15 years is its remarkable interannual variability, which is not yet fully explained. The largest contributions to interannual variability during the 1996–2001 period appear to be variations in emissions from rice agriculture and wetlands. A large CH_4 increase in 1998 has been attributed to climate related changes in wetland emissions and boreal forest fires in Siberia. [7.4]

Current atmospheric N_2O concentrations are about 18% higher than in the pre-industrial era and have been increasing almost linearly for the past few decades. The industrial-era increase of N_2O can be primarily attributed to human activities, particularly agriculture and associated microbial production of N_2O. Current estimates are that about 47% of total N_2O emissions are anthropogenic but individual source estimates remain subject to significant uncertainties. [2.3, 7.4]

TS.2.1.3 Radiative forcing due to halocarbons, stratospheric ozone, tropospheric ozone and other gases

The Montreal Protocol gases (primarily CFCs and HCFCs) as a group contributed 0.32 W m^{-2} to radiative forcing in 2004 with CFC-12 continuing to be the third most important long-lived radiative forcing agent. Although a slow decline has begun in the emissions and radiative forcing of some of these gases due to their phaseout under the Montreal Protocol, observations demonstrate that emissions of HCFCs and some CFCs are continuing. [2.3]

The concentrations of industrial fluorinated gases covered by the Kyoto Protocol (HFCs, PFCs, SF_6) are relatively small but are increasing. Their total radiative forcing in 2004 was 0.015 W m^{-2}. [2.3]

Stratospheric ozone is near its minimum level in the satellite observations era. Its radiative forcing is estimated to be –0.03 ± 0.07 W m^{-2}, weaker than quoted in TAR, with a medium level of scientific understanding. The total concentration of ozone-depleting substances has already peaked in the atmosphere and global stratospheric ozone may be beginning to show signs of recovery but is still ~4% below pre-1980 levels. The Antarctic ozone hole still forms every spring and is expected to continue to do so for many decades, contributing to radiative forcing. In addition to the chemical destruction of ozone, dynamical changes may have contributed to Northern Hemisphere midlatitude lower stratospheric ozone depletion. [2.3]

Tropospheric ozone radiative forcing is estimated to be 0.35 (+0.15/–0.1) W m^{-2} with a medium level of scientific understanding. The best-estimate of this radiative forcing has not changed since the TAR. Observations show that trends in tropospheric ozone during the last few decades vary in sign and magnitude at many locations, but there are indications of significant upward trends at low latitudes. Several new chemical transport model studies of the radiative forcing due to the increase in tropospheric ozone since preindustrial times exist and have increased complexity and comprehensiveness compared to models used in TAR. [2.3, 7.4]

Direct emission of water vapour by human activities makes a negligible contribution to radiative forcing. As global mean temperatures increase, water vapour concentrations increase. This represents a key feedback but not a forcing of climate change. In contrast, direct anthropogenic use of water does constitute a possible forcing, but corresponds to less than 1% of natural sources of water vapour. Water vapour has a short atmospheric lifetime (typically a few days), limiting the radiative forcing linked to direct emissions of water vapour by human activities. About 70% of the use of water for human activity is from irrigation. Over Asia where most of the irrigation takes place simulations suggest a change in the water vapour content in the lower troposphere by up to a 1%, resulting in an estimated radiative forcing of about 0.03 W m^{-2}, with a factor of three uncertainty and a very low level of scientific understanding. The direct injection of water vapour from fossil fuel combustion is significantly lower than the emission from agricultural activity. The radiative forcing from direct injection of water vapour by fossil fuel combustion specifically from aircraft has been estimated to be ~0.002 W m^{-2} [2.3, 2.5]

Stratospheric water vapour changes have likely contributed a positive radiative forcing of 0.07 ± 0.05 W m^{-2}, with a low level of scientific understanding. This estimate is based on chemical transport model studies of the effect of increases in methane, which leads to an indirect radiative forcing by increasing stratospheric water vapour. However, the level of scientific understanding is low because the causes of stratospheric water vapour change are only partially understood, especially in the lower stratosphere. [2.3]

TS.2.1.4 Sources and removals of halocarbons, stratospheric ozone, tropospheric ozone and other gases

Concentrations of CFC-11 and CFC-12 are expected to be affected by continuing leakage from "banks" such as foams, refrigeration, and air conditioning systems. New observations in polar firn cores since TAR have now extended the available time series information for many of these greenhouse gases. Ice core and in-situ data confirm that industrial sources are the cause of observed increases in CFCs and HCFCs. [2.3]

Improved measurements and related modelling have advanced the understanding of chemical precursors leading to the formation of tropospheric ozone, including carbon monoxide, nitrogen oxides (including sources and possible long-term trends in lightning) and formaldehyde. Overall, current tropospheric ozone models are successful in describing the principal features of the present-day global ozone distribution on the basis of underlying processes. New satellite and in-situ measurements provide important global constraints for models of tropospheric ozone and its precursors. However, much less confidence exists in the ability of these models to reproduce the changes in ozone associated with perturbations to emissions or climate, and in the simulation of observed long-term trends in ozone concentrations over the 20th century. [7.4]

Changes in tropospheric ozone are linked to air quality and climate change. A number of studies have shown that summer daytime ozone concentrations correlate strongly with temperature. This correlation appears to reflect contributions from temperature-dependent biogenic volatile organic carbon emissions, thermal decomposition of peroxyacetylnitrate, which acts as a reservoir for NO_x, and association of high temperatures with regional stagnation. Anomalously hot and stagnant conditions in the summer of 1988 were responsible for the highest surface-level ozone year on record in the northeastern United States. The summer heat wave in Europe in 2003 was also associated with exceptionally high local ozone at the surface. [Box 7.4]

TS.2.2 AEROSOLS

Direct aerosol radiative forcing is now considerably better understood than in TAR. A combined total direct aerosol radiative forcing is given as –0.5 ± 0.4 W m^{-2}, with a low level of scientific understanding. Satellite as well as surface based remote sensing and in-situ data and retrievals have developed considerably since the TAR (see Figure TS-4). Industrial aerosol consisting of a mixture of sulphates, organic and black carbon, nitrates, and industrial dust is clearly visible over many continental regions of the Northern Hemisphere. The total combined anthropogenic aerosol forcing is therefore now better quantified, although the specific contributions of the individual aerosol types remain less well constrained. The direct radiative forcing for individual species is estimated from models only to be: sulphate –0.4 ± 0.2 W m^{-2}, fossil-fuel organic carbon –0.1 ± 0.1 W m^{-2}, fossil-fuel black carbon +0.2 ± 0.1 W m^{-2}, biomass burning 0.0 ± 0.1 W m^{-2}, nitrate –0.1 ± 0.1 W m^{-2}, mineral dust –0.1 ± 0.2 W m^{-2}. It is very likely that radiative forcing from sulfate aerosol particles has decreased over about the past two decades, based both on ice core data and emissions estimates. [2.4, 6.6]

[INSERT FIGURE TS-4 HERE]

Significant changes in the estimates of the aerosol direct radiative forcing have occurred since the TAR for biomass burning, nitrate and mineral dust aerosols. For biomass burning aerosol the estimated direct radiative forcing is now revised from being strongly negative owing to better modelling of the effects of biomass burning on aerosol overlying clouds. For the first time, a radiative forcing for nitrate aerosol is given. For mineral dust, the range in the direct radiative forcing is reduced due to a reduction in the estimate of its anthropogenic fraction; a best estimate is given for the first time. [2.4]

The cloud-albedo radiative forcing due to aerosols (also referred to as first indirect or Twomey effect) is estimated to be -0.9 ± 0.5 W m^{-2}, with a very low (level of scientific understanding. The number of global model estimates of the albedo effect for warm (low-level) clouds has increased substantially since TAR. The estimate for this radiative forcing comes from multiple model studies incorporating more aerosol species and describing aerosol-cloud interaction processes in greater detail. Hence for the first time a best estimate is given for this radiative forcing. Observational studies, and model simulations constrained by observations, suggest a less negative estimate of the radiative forcing due to the albedo effect. Thus, the actual radiative forcing could be less negative than the model-based estimate above. However, there are only a few studies based on satellite datasets. This and other limitations, including uncertainties in the modelling of processes, introduce significant uncertainties in the radiative forcing estimate. Therefore the estimate of radiative forcing is associated with a very low level of scientific understanding. The magnitude of the overall indirect aerosol effect on clouds is estimated to be between –0.2 W m^{-2} and –2.0 W m^{-2}. The highest estimated value of the sum of these effects has been reduced from more than –4 W m^{-2} in TAR to –2.3 W m^{-2} because of improvements in cloud parameterizations and in understanding of aerosol microphysical processes. Ensemble-based studies that consider constraints from observations of climate change suggest that the net aerosol forcing over the 20th century is less than about –1.7 W m^{-2}. As in the TAR, other aerosol-cloud processes such as changes in cloud lifetime and semi-direct effects are accounted for in the evaluation of climate response rather than radiative forcing. Large uncertainties remain, because of interactions within the climate system and feedbacks on clouds, large-scale dynamics and the hydrological cycle, and therefore these processes have a very low level of scientific understanding. [2.4, 7.5]

TS.2.3 AVIATION CONTRAILS AND CIRRUS, LAND USE, AND OTHER EFFECTS

Aviation may alter cirrus clouds but the radiative forcing due to this is not quantified. Observational studies provide evidence that the net radiative forcing from spreading contrails and their effects on nearby cirrus cloudiness may be 2-10 times greater than the radiative forcing from persistent linear contrail cover. The global effect of aviation aerosol on background cloudiness remains unknown and a best estimate remains unavailable for the radiative forcing of total cloudiness changes (contrails, induced cirrus cloudiness, and aerosol effects) caused by subsonic aircraft operations. [2.6]

Persistent linear contrails from global aviation contribute a small radiative forcing of 0.01 W m^{-2}, with a factor of two uncertainty and a low level of scientific understanding. This best-estimate is a factor of 3 to 4 smaller than the estimate projected from the TAR for current global aviation operations. This difference results from new observations of contrail cover and reduced estimates of contrail optical depth. [2.6]

The impacts of land-use change on climate are expected to be locally significant in some regions, but are small at the global scale in comparison with greenhouse gas warming. Changes in the land surface (vegetation, soils, water) resulting from human activities can significantly affect local climate through shifts in radiation, cloudiness, and surface temperatures. Changes in vegetation cover can also have a substantial effect on surface energy and water balance at the regional scale. These effects have a very low level of scientific understanding. [7.2]

Human induced changes to land-cover have increased the global surface albedo. Other surface property changes lead to both radiative forcing and other physical alterations to the climate system. The combined radiative forcing from all surface-related mechanisms is estimated to be –0.1 ± 0.3 W m^{-2}, with a very low level of scientific understanding. Global anthropogenic land cover change since 1750 has consisted of more deforestation than tree planting, with most net deforestation occurring in temperate regions. The resulting increase in surface albedo has led to a global mean radiative forcing of –0.2 ± 0.3 W m^{-2}, with a low level of scientific understanding. Deposition of black carbon aerosols on snow decreases surface albedo and is estimated to give a radiative forcing of +0.1 W m^{-2}, with a factor of three uncertainty and a very low level of scientific understanding. [2.5]

The release of heat from anthropogenic energy production is significant at local scales in urban areas but not globally. [2.5]

TS.2.4 Radiative Forcing Due to Solar Activity and Volcanoes

Continuous monitoring of total solar irradiance now exists for 28 years and constrains the changes in solar forcing over that period to <0.05 W m^{-2}. An 11-year irradiance cycle of 0.08% (peak to peak) is well established, but no significant long-term trend has yet been detected. New data have more accurately quantified changes in solar spectral fluxes over a broad range of wavelengths in association with changing solar activity. In addition, improved calibrations are available using high quality overlapping instruments. Ultra-violet (UV) radiation at wavelengths below 310 nm contributes 15% of the total solar irradiance 11-year cycle. Total irradiance levels near the peak of cycle 23 (in 2001) were as high as during the two prior cycles, even though sunspot numbers were not, because both dark sunspots and bright faculae modulate the irradiance. The difficulty in estimating past variations in facular brightness contributes uncertainty to historical solar forcing estimates. Uncertainties also arise due to the lack of direct observations and understanding of solar variability mechanisms on long time scales, as well as the possibility of indirect effects. [2.7]

The estimated direct radiative forcing due to changes in the solar output since 1750 is 0.12 W m^{-2}, which is less than half of the estimate given in TAR; the estimate has a factor of two uncertainty and a low level of scientific understanding. The reduced radiative forcing estimate comes from a re-evaluation of the long-term change in solar irradiance since 1610 (the Maunder Minimum) based upon improved understanding of recent solar variability, its relationship to physical processes, and better information regarding the variability of sun-like stars. However, uncertainties remain large because of the lack of direct observations and understanding of solar variability mechanisms on long time scales. [2.7, 6.6]

A hypothesis that changes in cosmic rays associated with solar cycles could affect global average cloud cover, and thus climate, is not supported by available data. It has been suggested that galactic cosmic rays with sufficient energy to reach the troposphere could alter the population of cloud condensation nuclei and hence microphysical cloud properties (droplet number and concentration), inducing changes in cloud processes analogous to the indirect effect of tropospheric aerosols and representing an indirect solar forcing of climate. However, reported global correlations between the cosmic ray flux and mean cloud cover have diminished as more years of data became available. Thus, associations between galactic cosmic ray-induced changes in aerosol and cloud formation remain controversial. [2.7]

Explosive volcanic eruptions that greatly increase the concentrations of stratospheric sulphate aerosols can cool global mean climate for a few years. Volcanic aerosols perturb both the stratosphere and surface troposphere radiative energy budgets and climate in an episodic manner, and many past events are evident in ice core observations of sulphate as well as temperature records. There have been no intense volcanic events since the 1991 Pinatubo eruption. The global stratospheric aerosol concentrations are now at their lowest observed values since satellite measurements began in the late 1970s. The potential exists for volcanic eruptions much larger than the 1991 Pinatubo eruption, which would produce longer-term and larger radiative forcing of the climate system. [2.7, 6.4, 6.6, 9.3]

Improvements and extension of the time series of satellite measurements of volcanic aerosols in the stratosphere, and of solar variations, demonstrate that the net natural forcing (changes in volcanic plus solar) has been negative over the period of global scale observation (1978-present) [2.3, 2.7]

TS.2.5 Net Global Radiative Forcing, Global Mean Temperature Response, and Patterns of Forcing and Response

Humans activities since 1750 have very likely exerted a net warming influence on the Earth's climate. The global mean net radiative forcing due to the combined effects of LLGHGs and ozone is estimated to be 2.9 ± 0.3 W m^{-2}, which is the dominant radiative forcing term. Anthropogenic drivers that are less well understood (e.g., aerosol direct effects, cloud-albedo and land-use albedo) are likely to contribute a negative global mean radiative forcing. However, taking the sum of all anthropogenic drivers together, a net negative global-mean radiative forcing is very unlikely (see Figure TS-5). [2.3, 2.9, 9.2]

[INSERT FIGURE TS-5 HERE]

Although the spatial patterns of radiative forcing vary between different forcing agents, these perturbations in the energy balance drive many common climate processes and feedbacks. [9.2]

Considerable progress has been made in understanding the feedbacks between globally averaged temperature change and changes in water vapour and clouds, and have established that radiative forcing is directly related to a global mean climate response. Water vapour changes are a strong positive feedback that increases the global mean temperature change expected from changes in radiative forcing. New observational and modelling evidence strongly favours a combined water vapour – lapse rate[4] feedback of around the strength found in global climate models (GCMs) i.e. approximately 1 W m^{-2} per degree global temperature increase, or about a 40–50% amplification of global mean warming. Models have demonstrated an ability to simulate seasonal to inter-decadal humidity variations in the upper troposphere over land and ocean and have successfully simulated the observed surface temperature and humidity changes associated with volcanic eruptions. [2.3, 3.4, 8.6]

Confidence in the understanding that radiative forcing leads to climate responses is strengthened by the improved ability of current models to simulate past climate conditions. Many past large volcanic eruptions are evident as transitory cooling episodes in the instrumental and paleoclimate records, demonstrating that transient forcing leads to global-scale climate responses. The Last Glacial Maximum (LGM; the last 'ice age' about 21,000 years ago) and the mid-Holocene (6000 years ago) were different from the current climate not because of random variability, but because of altered seasonal and global forcing linked to known differences in the Earth's orbit (see Box TS.2.1). These led to important amplifying feedbacks, including changes in the ice sheets, in vegetation, and in greenhouse gas concentrations. For the mid-Holocene, coupled climate models are able to simulate a number of robust large-scale features of observed climate change, including mid-latitude warming and enhanced monsoons, with little change in global mean temperature (<0.4°C), consistent with our understanding of orbital forcing. There is also high confidence that changes in forcing linked to the Earth's orbit around the Sun were the principal driver for past ice ages. Biogeochemical and biogeophysical feedbacks amplified the response to orbital forcings. [6.2, 6.3, 6.4, 6.5, 9.3, 9.6]

[4] The rate at which air temperature decreases with altitude.

BOX TS.2.1: ORBITAL FORCING

It is well known from astronomical calculations that periodic changes in characteristics of the Earth's orbit around the Sun control the seasonal and latitudinal distribution of incoming solar radiation at the top of the atmosphere (hereafter called "insolation"). Past and future changes in insolation can be calculated over several millions of years with a high degree of confidence [6.4].

Although many aspects of orbital theory are well worked out, there is still no comprehensive mechanistic explanation for all of the observed climate and biogeochemical changes that take place over orbital time scales. Debate also continues on how the ice age cycles over the past 1 million years came to be dominated by ~100,000 year periods, given that the orbital forcing in this frequency band is relatively weak. [6.4, 6.5]

[INSERT BOX TS.2.1, FIGURE 1 HERE]

Precession refers to changes in the time of the year when the Earth is closest to the sun, with quasi-periodicities of about 19 and 23 kyr. As a result, changes in the position and duration of the seasons on the orbit strongly modulate the latitudinal and seasonal distribution of insolation. Seasonal changes of insolation are much larger than annual mean changes and can reach 60 W m^{-2} (Box TS.2.1, Figure 1).

The obliquity (tilt) of the Earth axis varies between about 22 and 24.5° with two neighbouring quasi-periodicities around 41 kyr. Changes in obliquity modulate seasonal contrasts as well as annual mean insolation changes with opposite effects in low versus high latitudes (and therefore no effect on global average of insolation) [6.4].

The eccentricity of the Earth's orbit around the Sun has longer quasi-periodicities at 400 kyr and around 100 kyr. Changes in eccentricity alone have limited impacts on insolation due to the resulting very small changes in Sun-Earth distance. However, changes in eccentricity interact with seasonal effects induced by obliquity and precession of the equinoxes. During periods of low eccentricity, such as ~400 kyr ago and during the next 100 kyr, seasonal insolation changes induced by precession are less strong than during periods of larger eccentricity (Box TS.2.1, Figure 1). [6.4]

The Milankovitch, or "orbital" theory of the ice ages is now well developed. Ice ages are generally triggered by changes in high northern hemisphere summer insolation minima, enabling winter snowfall to persist through the year and therefore accumulate to build northern hemisphere glacial ice sheets. Similarly, times with especially intense high northern hemisphere summer insolation, determined by orbital changes, are thought to trigger rapid deglaciations, associated climate change and sea level rise. These orbital forcings determine the pacing of climatic changes, while the large responses are to a large extent determined by strong feedback processes that amplify the orbital forcing. Over multi-millennial time scales, orbital forcing also exerts a major influence on key climate systems such as the Earth's major monsoons, the El Niño – Southern Oscillation, and global ocean circulation, as well as the greenhouse gas content of the atmosphere. There is no evidence that the current warming will be mitigated by a natural cooling trend towards glacial conditions. Understanding of the Earth's response to orbital forcing indicates that the earth would not naturally enter another ice age for at least 30,000 years. [6.4, 6.5]

Recent studies increase confidence that radiative forcing results in global mean temperature changes, and show that radiative forcing can be used to compare estimated global mean surface temperature changes for the range of human and natural influences on climate. The concept of efficacy measures the effectiveness of a given radiative forcing agent in causing a change in the global-mean surface temperature, compared to the same radiative forcing caused by a change in CO_2. Multiple model studies give improved confidence that efficacies for realistic anthropogenic and natural radiative forcings lie within the range 0.75–1.25. Thus there is high confidence that global-mean radiative forcing gives a quantitative comparative estimate of the equilibrium global mean surface temperature change for realistic climate change mechanisms. [2.9]

The Global Warming Potential (GWP) remains a useful metric for comparing the potential climate impact of the emissions of different LLGHGs (see Table TS-2). GWPs compare the integrated radiative forcing over a specified period (e.g., 100 years) from a unit mass pulse emission and are a way of comparing the potential climate change associated with emissions of different agents. There are well-documented shortcomings of the GWP concept, particularly in using it to assess the impact of short-lived species. The future climate impact of current emissions of LLGHGs considered under the Kyoto Protocol can be compared through consideration of so-called CO_2-equivalent emissions, i.e. emissions multiplied by the GWP_{100} for the non-CO_2 gases (see Figure TS-6). The sum of methane, nitrous oxide, and other gases are estimated to contribute about half as much total GWP-weighted emission in the year 2000 as CO_2. Uncertainties in the estimates of the equivalent CO_2 emissions originate both from uncertainties in lifetimes and optical properties (through the GWP values) as well as uncertainties in the current global emissions. Compounds with longer lifetimes make different contributions to century-scale future climate change as estimated in this figure compared to the radiative forcing bar-chart diagram (see Figure TS-6). [2.10]

[INSERT FIGURE TS-6 HERE]

[INSERT TABLE TS-2 HERE]

For the magnitude and range of realistic radiative forcings considered, evidence suggests an approximately linear relationship between global mean radiative forcing and global mean surface temperature response. However, the spatial signature of a climate model's response is not generally expected to correspond to that of the forcing. This comes about because the climate system can move heat around and because feedbacks vary spatially. For example, sea ice albedo feedbacks tend to enhance the high-latitude response. Spatial patterns of response are affected by factors including differences in thermal inertia between land and sea areas, and the lifetimes of the various forcing agents. [2.8, 9.2]

The pattern of response to a radiative forcing can be altered substantially if its structure is favourable for affecting a particular aspect of the atmospheric structure or circulation. Modelling studies and data comparisons suggest that mid- to high-latitude circulation patterns are likely to be affected by some forcings such as volcanic eruptions, which have been linked to changes in the Northern Annular Mode (NAM) and North Atlantic Oscillation (NAO) (see Box TS.3.1). Simulations also suggest that absorbing aerosols, particularly black carbon, can reduce the solar radiation reaching the surface and can warm the atmosphere on regional scales, affecting the vertical temperature profile and the large-scale atmospheric circulation [2.8, 7.5, 9.2]

The spatial patterns of radiative forcings for ozone, aerosol direct effects, aerosol-cloud interactions and land-use have considerable uncertainties. This is in contrast to the relatively high confidence in the spatial pattern of radiative forcing for the LLGHGs. The positive net radiative forcing in the Southern Hemisphere very likely exceeds that in the Northern Hemisphere as a result of the negative aerosol radiative forcing, which is concentrated more in the Northern Hemisphere [2.9]

TS 2.6 SURFACE FORCING AND THE HYDROLOGIC CYCLE

Observations and models indicate that changes in the radiative flux at the Earth's surface affect the surface heat and moisture budgets, thereby linking to the hydrologic cycle. While evidence suggests that global mean temperature changes are consistent across the range of radiative forcing agents, recent studies indicate that some forcing agents can influence the hydrologic cycle differently than others. In particular, changes in total aerosols have very likely reduced the global temperature increase during the 20th century, but may also have affected precipitation and other aspects of the hydrologic cycle more strongly than other anthropogenic forcing agents. The instantaneous radiative flux change at the surface (hereafter called "surface forcing") is a useful diagnostic tool for understanding changes in the heat and moisture surface budgets and the accompanying climate change. However, unlike radiative forcing, it cannot be used to quantitatively compare the effects of different agents on the equilibrium global-mean surface temperature change. Energy deposited at the surface can directly affect evaporation. The net radiative forcing and surface forcing display different equator-to-pole gradients and Northern-to-Southern Hemisphere ratios than the total positive radiative forcing, which could affect precipitation patterns. [2.8, 7.2, 7.5, 9.5]

Table TS-2. (GWP Table, Table 2.14) Lifetimes, radiative efficiencies, and direct (except for methane) global warming potentials (GWP) relative to CO_2. For ozone depleting substances and their replacements data are taken from IPCC/TEAP (2005) unless otherwise indicated.

Industrial Designation or Common Name	Chemical Formula	Other Name	Lifetime (years)	Radiative Efficiency ($W\ m^{-2}\ ppb^{-1}$)	Global Warming Potential for Given Time Horizon (years)			
					SAR (100)	20	100	500
Carbon dioxide	CO_2		See below[a]	See below[b]	1	1	1	1
Methane[c]	CH_4		10.8[c]	3.7×10^{-4}	21	67	23	6.9
Nitrous oxide	N_2O		114[c]	3.1×10^{-3}	310	291	298	153
Substances controlled by the Montreal Protocol								
CFC-11	CCl_3F	Trichlorofluoromethane	45	0.25	3800	6700	4750	1620
CFC-12	CCl_2F_2	Dichlorodifluoromethane	100	0.32	8100	11000	10800	5200
CFC-113	CCl_2FCClF_2	1,1,2-Trichlorotrifluoroethane	85	0.3	4800	6540	6130	2700
CFC-114	$CClF_2CClF_2$	Dichlorotetrafluoroethane	300	0.31		8040	10000	8700
CFC-115	$CClF_2CF_3$	Monochloropentafluoroethane	1700	0.18		5310	7370	10000
Halon-1301	$CBrF_3$	Bromotrifluoromethane	65	0.32	5400	8480	7140	2760
Halon-1211	$CBrClF_2$	Bromochlorodifluoromethane	16	0.3		4750	1890	575
Halon-2402	$CBrF_2CBrF_2$	1,2-Dibromotettrafluoroethane	20	0.33		3680	1640	500
Carbon tetrachloride	CCl_4	(Halon-104)	26	0.13	1400	2700	1400	435
Methyl bromide	CH_3Br	(Halon-1001)	0.7	0.01		17	5	1
Bromochloromethane	CH_2BrCl	(Halon-1011)	0.37					
Methyl chloroform	CH_3CCl_3	1,1,1-Trichloroethane	5	0.06		510	146	45
HCFC-22	$CHClF_2$	Chlorodifluoromethane	12	0.2	1500	5200	1800	550
HCFC-123	$CHCl_2CF_3$	Dichlorotrifluoroethane	1.3	0.14	90	270	77	24
HCFC-124	$CHClFCF_3$	Chlorotetrafluoroethane	5.8	0.22	470	2070	610	185
HCFC-141b	CH_3CCl_2F	Dichlorofluoroethane	9.3	0.14		2250	730	220
HCFC-142b	CH_3CClF_2	Chlorodifluoroethane	17.9	0.2	1800	5500	2300	705
HCFC-225ca	$CHCl_2CF_2CF_3$	Dichloropentafluoropropane	1.9	0.2		430	120	37
HCFC-225cb	$CHClFCF_2CClF_2$	Dichloropentafluoropropane	5.8	0.32		2030	600	180
Hydrofluorocarbons								
HFC-23	CHF_3	Trifluoromethane	270	0.19	11700	12000	14800	12200
HFC-32	CH_2F_2	Difluoromethane	4.9	0.11	650	2330	675	205
HFC-125	CHF_2CF_3	Pentafluoroethane	29	0.23	2800	6350	3500	1100
HFC-134a	CH_2FCF_3	1,1,1,2-Tetrafluoroethane	14	0.16	1300	3830	1430	435
HFC-143a	CH_3CF_3	1,1,1-Trifluoroethane	52	0.13	3800	5890	4470	1590

HFC-152a	CH_3CHF_2	1,1-Difluoroethane	1.4	0.09	140	437	124	38
HFC-227ea	CF_3CHFCF_3	1,1,1,2,3,3,3-Heptafluoropropane	34.2	0.26	2900	5300	3200	1000
HFC-236fa	$CF_3CH_2CF_3$	1,1,1,3,3,3-Hexafluoropropane	240	0.28	6300	8100	9800	7660
HFC-245fa	$CHF_2CH_2CF_3$	1,1,1,3,3-Pentafluoropropane	7.6	0.28		3400	1030	310
HFC-365mfc	$CH_3CF_2CH_2CF_3$	1,1,1,3,3-Pentafluorobutane	8.6	0.21		2500	790	240
HFC-43-10mee	$CF_3CHFCHFCF_2CF_3$	1,1,1,2,2,3,4,5,5,5-Decafluoropentane	15.9	0.4	1300	4100	1640	500
Perfluorocarbons								
	SF_6	Sulfur hexafluoride	3200	0.52	23900	16300	22800	32600
	NF_3	Nitrogen trifluoride	740	0.13		8300	11200	13200
PFC-14	CF_4	Carbon tetrafluoride	50000	0.10	6500	4200	5900	8950
PFC-116	C_2F_6	Perfluoroethane	10000	0.26	9200	8600	12200	18200
PFC-218	C_3F_8	Perfluoropropane	2600	0.26	7000	6300	8800	12500
PFC-318	$c\text{-}C_4F_8$	Perfluorocyclobutane	3200	0.32	8700	7300	10300	14700
PFC-3-1-10	C_4F_{10}	Perfluorobutane	2600	0.33	7000	6300	8900	12500
	C_5F_{12}	Perflouropentane	4100	0.41		6500	9200	13300
PFC-5-1-14	C_6F_{14}	Perfluorohexane	3200	0.49	7400	6600	9300	13300
	$C_{10}F_{18}$	Perfluorodecalin	1000[d]	0.56		5500	7500[d]	9500
	SF_5CF_3	Trifluoromethyl sulfur pentafluoride	1000[e]	0.57		13000	18000[d]	23000
Fluorinated ethers								
HFE-125	CF_3OCHF_2		150	0.44		13900	15400	9300
HFE-134	CHF_2OCHF_2		26.2	0.45		12200	6360	1980
HFE-143a	CH_3OCF_3		4.4	0.27		3420	1000	305
HCFE-235da	$CF_3CHClOCHF_2$		2.6	0.38		1230	350	106
HFE-245cb2	$CF_3CF_2OCH_3$		4.3	0.32		2080	597	182
HFE-245fa2	$CF_3CH_2OCHF_2$		4.4	0.31		2060	592	180
HFE-254cb2	$CHF_2CF_2OCH_3$		0.22	0.28		107	30	9
HFE-347mcc3	$CF_3CF_2CF_2OCH_3$		4.5	0.34		1730	500	150
HFE-347pcf2	$CF_3CH_2OCF_2CHF_2$		7.1	0.25		1900	580	175
HFE-356pcf3	$CHF_2CF_2CH_2OCHF_2$		3.2	0.39		1570	450	140
HFE-449sl	$CH_3O(CF_2)_3CF_3$	(HFE-7100)	5	0.31		1390	404	123
HFE-569sf2	$CH_3CH_2O(CF_2)_3CF_3$	(HFE-7200)	0.77	0.3		200	57	17
H-Galden 1040x	$CHF_2OCF_2OC_2F_4OCHF_2$		6.3	1.37		6300	1900	570
HG-10	$CHF_2OCF_2OCHF_2$		12.1	0.66		8000	2800	860
HG-01	$CHF_2OCF_2CF_2OCHF_2$		6.2	0.87		5100	1500	460
Hydrocarbons and other compounds								

- Direct Effects							
Dimethylether	CH_3OCH_3		0.015	0.02	1	1	<<1
Methylenechloride	CH_2Cl_2	(Freon-40) Dichloromethane	0.38	0.03	31	8.7	2.7
Methyl chloride	CH_3Cl	(Freon-30) Chloromethane	1.3	0.01	59	17	5.1

Notes:
(a) The CO_2 response function used in this report is based on the revised version of the Bern Carbon cycle model used in Chapter 8 of this report using a background CO_2 concentrations 378 ppm. The decay of a pulse of CO_2 with time *t* is given by

$a_0 + \sum_{i=1}^{3} a_i \cdot e^{-t/\tau_i}$, where a_0=0.217, a_1=0.259, a_2=0.338, a_3=0.186, τ_1=172.9 years, τ_2=18.51 years, and τ_3=1.186 years.

(b) The radiative forcing of CO_2 is calculated by the expression RF=α ln(C/C_0), where α=5.35 (IPCC, 2001) and C_0 = 378 ppm. The radiative efficiency of CO_2 is the increase in forcing for a 1 ppb increase in abundance, or 1.4 x 10^{-5} W m^{-2} ppb^{-1}.
(c) The adjustment time for methane has been reduced from 12 years since the TAR (cf. Chapter 7, Section 7.4). The GWP for methane includes indirect effects from enhancements of ozone and stratospheric water vapour (see Chapter 2, Section 2.10)
(d) The lifetime is very uncertain, the assumed lifetime of 1000 years is a lower limit.
(e) The assumed lifetime of 1000 years is a lower limit.

TS.3 OBSERVATIONS OF CHANGES IN CLIMATE

The global mean of the radiative forcing and of the climatic response is a spatial average over a system that varies on all space and time-scales. The atmosphere and ocean are fluids that transport energy from the tropics (with its excess in radiation) to the higher latitudes (with its deficit). Regional climate depends on the characteristics of the atmospheric and oceanic circulation, the cryosphere, and the land surface and the interactions among them. The global mean trends of temperature and precipitation are expected to be linked to global mean net radiative forcing. However the local response is complex and may reflect coupled changes in atmosphere-ocean circulation and variability, such as El Niño events.

This assessment not only documents changes in the atmosphere, the ocean, and the cryosphere, but also changes in phenomena such as atmospheric circulation changes, in order to increase understanding of the character of climate change at global and regional scales.

TS.3.1 ATMOSPHERIC CHANGES: INSTRUMENTAL RECORD

Documentation of climate change includes analysis of global and hemispheric means, changes over land and ocean, and distributions of trends in latitude, longitude, and altitude. Improvements in observations and their calibration, more detailed analysis of methods, and extended time series now allow more in-depth analyses of atmospheric changes including atmospheric temperature, precipitation, humidity, wind, and circulation than previously possible. Extremes of climate are a key expression of climate variability, and this assessment also includes new data that permits improved insights into the changes in many types of extreme events including heat waves, droughts, heavy precipitation, and tropical cyclones (i.e., hurricanes and typhoons). [3.1, 3.2]

Further, advances have occurred in understanding how a number of seasonal and long-term anomalies can be described by patterns of variability. These patterns arise from the differential effects on the atmosphere of land and ocean, mountains, and anomalous heating. The response is often felt in regions far removed from the anomalous heating through atmospheric teleconnections, associated with large-scale waves in the atmosphere. Changes in temperature and precipitation anomalies associated with the dominant patterns are essential to understanding many regional climate anomalies and why these may differ from global means. Changes in storm tracks, the jet streams, regions of preferred blocking anticyclones, and changes in monsoons can occur in conjunction with these preferred patterns. [3.5]

TS.3.1.1 Globally averaged temperatures

2005 and 1998 were the warmest two years on record. Five of the six warmest years have occurred in the last five years (2001-2005). [update to include 2006 before the final WG1 plenary in 2007] [3.2]

The global average surface temperature has increased since 1850. The linear warming trend over the 20th century was 0.6 ± 0.2°C in the TAR. For the period from 1901–2005 it is 0.65 ± 0.2 due to additional warm years. The record shows substantial variability, and a linear trend is not a good fit to the data. Most of the warming occurred from 1910–1945 (0.14°C decade^{-1}) and 1979–2005 (0.17°C decade^{-1}). Three different global estimates all show consistent warming trends. Agreement is also consistent between the datasets in their separate land and ocean domains, and between SST and night-time marine air temperature. [update to include 2006 before the final WG1 plenary in 2007] (See Figure TS-7) [3.2]

[INSERT FIGURE TS-7 HERE]

Recent studies show that effects of urbanization and land use change on the land-based temperature record (since 1950) are negligible as far as hemispheric- and continental-scale averages are concerned. Increasing evidence suggests that while urban heat island effects may extend beyond local temperatures to changes in precipitation, cloud, and DTR, these effects are negligible as far as hemispheric and continental-scale averages are concerned. The very real but local effects of urban areas are accounted for in the land temperature datasets used are not relevant to the widespread oceanic warming that has been observed. [3.2]

The global average diurnal temperature range has changed. The globally averaged (71% of the land surface where data are available) diurnal temperature range (DTR) decreased 0.07ºC decade^{-1} over the period 1950-2004. However, there was very little change over the 1979–2004 period. [3.2]

New analyses of radiosonde and satellite measurements of lower-tropospheric temperature now show warming rates that are quantitatively consistent within error bars with the surface temperature record over the periods 1958–2005 and 1979–2005, respectively (see Figure TS-8). The radiosonde record is markedly less spatially complete than the surface and increasing evidence suggests a number of these records are unreliable, especially in the tropics. While there remain disparities among different tropospheric temperature trends estimated from satellite microwave sounder unit (MSU) and advanced MSU (AMSU) measurements since 1979, and all likely still contain residual errors, trend estimates have been substantially improved (and dataset differences reduced) since the TAR through adjustments for issues of changing satellites, orbit decay, and drift in local crossing time (diurnal cycle effects). The range (due to different datasets) of global surface warming since 1979 is 0.16 to 0.18 compared to 0.12 to 0.19ºC decade^{-1} for MSU estimates of tropospheric temperatures. It is likely that there is increased warming with altitude from the surface through much of the troposphere in the tropics, pronounced cooling in the stratosphere, and a trend towards a higher tropopause. [3.4]

[INSERT FIGURE TS-8 HERE]

Stratospheric temperature estimates from adjusted radiosondes, satellites, and reanalyses are all in quantitative agreement, with a cooling of between 0.3 and 0.6°C decade^{-1} since 1979 (see Figure TS-8). Longer radiosonde records (back to 1958) also indicate stratospheric cooling but are subject to substantial instrumental uncertainties. The rate of cooling has been significantly greater since 1979 than between 1958 and 1978. [3.4]

TS.3.1.2 Spatial distribution of changes in temperature, circulation and related variables

Observations since the TAR demonstrate that land regions have warmed at a faster rate than the oceans in both hemispheres. Longer records now available show significantly faster rates of warming over land than ocean in the past two decades (about 0.25 versus 0.13°C decade^{-1}). [3.2]

Some features seen in the globally averaged surface temperature changes are seen in the zonally averaged temperatures at all latitudes. The warming in the last 30 years is widespread over the globe, and is a maximum at higher northern latitudes. The warming from 1920 to 1945 and the cooling from 1946 to 1978 did not occur at all latitudes. Arctic temperatures have been warming on average since the 1960s, and 2005 was the warmest Arctic year. However, Arctic temperatures are variable, and a warm period was also observed from 1920–1945. [3.2]

There is increasing evidence for changes in the large-scale atmospheric circulation, such as a poleward shift and strengthening of the westerly winds. Regional climate trends over a few decades can be very different from the global average, reflecting changes in the circulations and interactions of the atmosphere and ocean and the other components of the climate system. Increasing westerly winds have occurred in both hemispheres in most seasons from 1979 to the late 1990s, and poleward displacements of corresponding Atlantic and southern polar front jet streams have been documented. The increasing strength of the westerlies in the NH changes the flow from oceans to continents, and is a major factor in the observed wintertime changes in storm tracks and related patterns of precipitation and temperature trends at mid- and high-latitudes. These changes have been accompanied by a tendency toward stronger wintertime polar vortices throughout the troposphere and lower stratosphere. The greatest warming has occurred in the northern hemisphere winter (DJF) and spring (MAM). [3.2, 3.5]

Many regional climate changes can be described in terms of preferred patterns of climate variability and therefore as changes in the occurrence of values of the indices of these. For many years the importance on all time-scales of fluctuations in the westerlies and the storm-track in the North Atlantic have been noted and described by the NAO (see Box TS.3.1). The importance of fluctuations in the zonally averaged westerlies in the two hemispheres has more recently been described by their respective " annular modes" NAM and SAM (see Figure TS-9). The observed changes can be expressed as positive biases in the occurrence of these preferred patterns. The increased middle latitude westerlies in the North Atlantic can be largely viewed as either NAO or NAM changes. In the SH, SAM changes are identified with contrasting trends of the strong warming over the Antarctic Peninsula, and cooling over much of continental Antarctica. Multi-decadal variability is also evident in the Atlantic, both in the atmosphere and ocean. Changes have also been observed in ocean-atmosphere interactions in the Pacific. ENSO is the dominant mode of global-scale variability on interannual time scales although there have been times when it is less apparent. The 1976–1977 climate shift, related to the phase change in the PDO toward more El Niños and changes in the evolution of ENSO, has affected many areas, including most tropical monsoons. For instance, over North America, ENSO and PNA teleconnection-related changes appear to have led to contrasting changes across the continent, as the west has warmed more than the east, while the latter has become cloudier and wetter. There is substantial low-frequency atmospheric variability in the Pacific sector over the 20th century, with extended periods of weakened (1900–1924; 1947–1976) as well as strengthened circulation (1925–1946; 1977–2003). [3.2, 3.5, 3.6]

[INSERT FIGURE TS-9 HERE]

BOX TS.3.1: PATTERNS (MODES) OF CLIMATE VARIABILITY

Analysis of atmospheric/climate variability has shown that a significant component of it can be described in terms of fluctuations in the amplitude and sign of indices of a relatively small number of preferred patterns of variability. Some of the best known of these are:

- El Niño-Southern Oscillation (ENSO), a coupled fluctuation in the atmosphere and the equatorial Pacific Ocean, with preferred times scales of 2 to about 7 years. ENSO is often measured by the surface pressure anomaly difference between Darwin and Tahiti and the sea surface temperatures in the central and eastern equatorial Pacific. ENSO has global teleconnections with fluctuations elsewhere.
- North Atlantic Oscillation (NAO), a measure of the strength of the Icelandic Low and the Azores high, and also of the westerly winds between them, mainly in winter. The NAO has associated fluctuations in the storm-track, temperature and precipitation from the North Atlantic into Eurasia.
- Northern Annular Mode (NAM), a winter-time fluctuation in the amplitude of a pattern characterised by low surface pressure in the Arctic and strong middle latitude westerlies. The NAM has links with the northern polar vortex into the stratosphere. Its pattern has a bias to the North Atlantic and has a large correlation with the NAO.
- Southern Annular Mode (SAM), the fluctuation of a pattern like the NAM but in the Southern Hemisphere.
- Pacific North American (PNA) pattern, an atmospheric large-scale wave pattern featuring a sequence of tropospheric high and low pressure anomalies stretching from the subtropical west Pacific to the east coast of North America.
- Pacific Decadal Oscillation (PDO), a measure of the sea surface temperatures in the North Pacific that has a very strong correlation with the North Pacific Index (NPI) measure of the depth of the Aleutian Low. However, it has a signature throughout much of the Pacific.

The extent to which all these preferred patterns of variability can be considered to be true modes of the climate system is a topic of active research. However there is evidence that their existence can lead to larger amplitude responses to forcing than would otherwise be expected. In particular, a number of the observed 20th century climate changes can be viewed in terms of changes in them. It is therefore important that climate models should be able to simulate them well [see Box TS.4.1] and to consider whether changes in these patterns may not be entirely linked to internal variability but rather may be altered by anthropogenic climate changes. [8.4]

Wind and significant wave height analysis provide new supporting evidence for an increase in extratropical storm activity in the Northern Hemisphere in recent decades. [3.5]

Changes in extremes of temperature are consistent with warming. A widespread reduction in the number of frost days in mid-latitude regions, an increase in the number of warm extremes and a reduction in the number of daily cold extremes are observed (see Box TS.3.4). The most marked changes are for cold (lowest 10%) nights, which have declined over the 1951–2003 period for 76% of the land regions studied. Warm (highest 10%) nights have increased across 72% of the same regions [3.8]

Heat waves have increased in duration beginning in the latter half of the 20th century. Widespread increases have been documented in the occurrence of warm nights. The record-breaking heat wave over western and central Europe in the summer of 2003 is an example of an exceptional recent extreme. This summer (JJA) was the warmest since comparable instrumental records began around 1780 (1.4ºC above the previous warmest in 1807). Spring drying of the land surface over Europe was an important factor in the occurrence of the extreme 2003 temperatures. Evidence suggests that heat waves have also increased in frequency and duration in other locations. The very strong correlation between observed dryness and high temperatures over land in the summer and the tropics highlights the important role moisture plays in moderating climate. [3.8]

TS.3.1.3 Changes in the water cycle: water vapour, clouds, precipitation, and tropical storms

Tropospheric water vapour is increasing. Averaged surface specific humidity[5] has generally increased after 1976 in close association with higher temperatures over both land and ocean[6]. Total column water vapour has increased over the global oceans by 1.2 ± 0.3% (95% confidence limits) from 1988 to 2004, consistent in pattern and amount with changes in SST and a fairly constant relative humidity (see Figure TS-10). Strong correlations with SST suggest that water vapour has increased by 4% over the global oceans since 1970. Increases in water vapour have also been documented in the global upper troposphere from 1982–2004 using new information from satellites. [3.4]

[INSERT FIGURE TS-10 HERE]

Cloud changes are dominated by ENSO. Widespread (but not ubiquitous) decreases in continental DTR since the 1950s coincide with increases in cloud amounts. Total and low-level cloud changes over the ocean disagree between surface and satellite observations. However, radiation changes at the top-of-the atmosphere from the 1980s to 1990s, possibly related in part to the ENSO phenomenon, appear to be associated with reductions in tropical upper-level cloud cover, and are expected to be linked to changes in the energy budget at the surface and in observed ocean heat content. [3.4]

"Global dimming" is neither global in extent nor has it continued after 1990. Reported decreases in solar radiation at the Earth's surface from 1970 to 1990 have an urban bias and have reversed in sign. An increasing aerosol load due to human activities decreases regional air quality and the amount of solar radiation reaching the earth's surface. In some areas, such as Eastern Europe, recent observations of a reversal in sign of this effect link changes in solar radiation to concurrent air quality improvements. Although records are sparse, pan evaporation is estimated to have decreased in many places due to decreases in surface radiation associated with increases in clouds, changes in cloud properties, and/or increases in air pollution (aerosol) in different regions, especially from 1970 to 1990. However, observations in many areas suggest that actual evapotranspiration inferred from surface water balance exhibits an increase in association with enhanced soil wetness from increased precipitation, as the actual evapotranspiration becomes closer to the potential evaporation measured by the pans. Hence there is a trade-off in evapotranspiration between less solar radiation and increased wetness, with the latter generally dominating. [3.4, 7.5, 9.2]

Patterns of precipitation change are emerging, with some latitudes displaying decreases in total precipitation while others have increased. Zonal averages indicate less precipitation in the northern subtropics after about 1970, and increases at higher latitudes (Figure TS-11 upper panel). The geographical

[5] *Specific* or *absolute humidity* is the mass of water vapor per unit mass of air, including the water vapor, and usually expressed as grams of water vapor per kilogram of air. Warmer air has a higher possible maximum absolute humidity than cooler air, because condensation limits the maximum humidity that can be contained in the air. Relative humidity is a measure of the amount of water in air compared with the amount of water the air can contain before condensation occurs.

[6] A non-linear linkage between temperature and water vapor is established by the physics of the Clausius-Clapeyron relationship and nearly constant relative humidity (but changing absolute humidity) with increasing temperatures.

distribution of the trends over the period 1901–2005 (Figure TS-11, lower panel) indicates the spatial variability, although there are concerns about the length and consistency of records. There has been a reduction in precipitation in northern subtropical Africa. Precipitation generally decreased in the deep tropics from 10°N to 10°S, especially after 1976/1977. [3.3]

[INSERT FIGURE TS-11 HERE]

Substantial increases have been observed in heavy precipitation events. It is likely that there have been increases in the number of heavy precipitation events (e.g., 95th percentile) in many land regions, even those where there has been a reduction in total precipitation amount. Increases have also been reported for rarer precipitation events (1 in 50 year return period), but only a few regions have sufficient data to assess such trends reliably (see Figure TS-12). [3.8]

[INSERT FIGURE TS-12 HERE]

There is no clear trend in the total numbers of tropical cyclones. More storms in one ocean basin are often compensated by fewer in another. The numbers and proportion of hurricanes reaching categories 4 and 5 globally have increased since 1970, while total number of cyclones and cyclone days decreased slightly in most basins. The largest increase was in the North Pacific, Indian and Southwest Pacific oceans. Data quality and coverage issues, particularly prior to the satellite era, represent obstacles to analysis on longer time scales. Variations in the total numbers of tropical cyclones, hurricanes and typhoons are dominated by ENSO and decadal variability, which result in a redistribution of tropical storm numbers and their tracks. However, numbers of hurricanes in the North Atlantic have been above normal (based on 1981–2000) in 9 of the last 11 years, culminating in the record-breaking 2005 season. [3.8]

There is evidence that the intensity and duration of tropical cyclones has increased since the 1970s although there are concerns about the quality of the historical data. Globally, estimates of the potential destructiveness of hurricanes show a substantial upward trend since the mid-1970s, with a trend toward longer lifetimes and greater storm intensity (see Figure TS-13). Trends are also apparent in SSTs and other critical variables that influence tropical storm development. [3.8]

[INSERT FIGURE TS-13 HERE]

Droughts have been widespread in various parts of the world since the 1970s. Direct relationships to global warming have been inferred through the nature of high temperatures and heat waves accompanying recent droughts. [3.3]

TS.3.2 CHANGES IN THE CRYOSPHERE: INSTRUMENTAL RECORD

Presently, ice permanently covers 10% of the land surface, with only a tiny fraction occurring outside Antarctica and Greenland. Ice also covers approximately 6.5% of the oceans in the annual mean. In mid-winter, snow covers approximately 49% of the land surface in the Northern Hemisphere (NH). An important property of snow and ice is its high surface albedo. Up to 90% of the incident solar radiation is reflected by snow and ice surfaces, while only about 10% is reflected by the open ocean or forested lands. In addition, snow and ice are effective insulators. Seasonally frozen ground is more extensive than snow cover, and its presence is also important for energy and moisture fluxes. Therefore, frozen surfaces play important roles in the energy balance of the Earth. [4.1]

In addition, the cryosphere stores about 75% of the world's fresh water. The volume of the Greenland and Antarctic ice sheets are equivalent to approximately 7 m and 57 m of sea level rise, respectively. New analyses since the TAR have improved the understanding of how changes of the ice mass on land contributed to recent changes of the sea level. On a regional scale, mountain snowpack, glaciers and small ice caps play a crucial role in fresh water availability. Since the change from ice to liquid water occurs at specific temperatures, ice is a component of the climate system that could be subject to abrupt change following small changes in temperature. Changes in ice in some parts of the globe include shrinkage of mountain glacier volume, earlier springtime melt of snow cover, reductions in Arctic sea ice extent, coastal thinning of the large ice sheets exceeding inland thickening from increased snowfall, and reductions in seasonally frozen ground, river and lake ice cover. [4.1]

Snow cover has decreased in most regions, especially in spring. Average April Northern Hemisphere snow cover observed by satellite decreased during the 1966–2004 period by 0.5×10^6 km^2 per decade, or about 5% in total (see Figure TS-14). Average Northern Hemisphere snow cover decreased in every month but November and December. In the Southern Hemisphere, the handful of long records or proxies mostly shows either decreases or no changes in the past 40+ years. The decrease has been especially prominent in late winter and spring, with little change in fall or early winter, and has occurred in many places despite increases in extra-tropical precipitation. [4.2]

[INSERT FIGURE TS-14 HERE]

Decreases in snowpack have been documented in several regions worldwide based upon annual time series of mountain snow water equivalent and snow depth. Mountain snow can be sensitive to small changes in temperature, particularly in temperate climatic zones where the transition from rain to snow is often closely associated with altitude. Mountain snow water equivalent has declined since 1950 at 75% of the stations monitored in western North America. Mountain snow depth has also declined in the Alps and in southeastern Australia. Direct observations of snow depth are too limited to determine changes in the Andes, but temperature measurements suggest that the altitude where snow occurs (the snowline) has probably risen in mountainous regions of South America. [4.2]

Permafrost and seasonally frozen ground display large changes in recent decades. Changes in permafrost can affect river runoff, water supply, carbon exchange, rock falls, and cause damage to infrastructure. Permafrost temperature has increased by up to 3°C since the 1980s. The permafrost base is thawing at a rate ranging from 0.02 m/year in Alaska to 0.4 m/year on the Tibetan Plateau. Permafrost warming is also observed with variable magnitude but a consistent trend in the Canadian Arctic, Siberia, Tibetan Plateau, and Europe. Permafrost boundaries have moved northwards in Canada and upwards on the Tibetan Plateau, consistent with warming. [4.7]

The maximum area covered by seasonally frozen ground decreased by about 7% in the NH over the latter half of the 20th century. Its maximum depth has decreased by about 0.3 m in Eurasia since the mid-20th century. Evidence from satellite passive microwave remote sensing indicates that the onset dates of thaw in spring and freeze in autumn advanced five to seven days in Eurasia from 1988–2002, leading to an earlier start to the growing season but no change in its length. [4.7]

On average, the general trend of river and lake ice over the past 150 years indicates that the freeze-up date has become later at a rate of 5.8 ± 1.9 days per century, while the break-up date has occurred earlier, at a rate of 6.5 ± 1.4 days per century. However, considerable spatial variability is also observed. [4.3]

Annually averaged Arctic sea ice extent has shrunk by about 2.7% ± 0.7% *per decade since 1978 based upon satellite observations (see Figure TS-15).* The decline for summertime extent is larger than for wintertime, with the summer minimum declining at a rate of about 7.4% ± 2.9% per decade. Other data indicate that the summer decline began around 1970. The least summer Arctic sea ice was observed in 2005. Similar observations in the Antarctic reveal larger inter-annual variability but no consistent trends during the period of satellite observations. In contrast to changes in continental ice such as ice sheets and glaciers, changes in sea ice do not contribute to sea level change because this ice is already floating, but can contribute to salinity changes through input of freshwater. [4.4]

[INSERT FIGURE TS-15 HERE]

Ice thickness data from submarines indicate a reduction in average Arctic sea-ice thickness of about 1m from 1987 to 1997. Ice thickness information in the Antarctic is too sparse to make inferences regarding trends. [4.4]

During the 20th century, glaciers and ice caps have experienced widespread mass losses with strongest retreats in the 1930s and 1940s and after 1990. Mass loss of glaciers and ice caps (excluding those around *the ice sheets of Greenland and Antarctica), is estimated to have contributed significantly to sea level rise: by 0.51 ± 0.32 mm per year in sea level equivalent (SLE) between 1961 and 2003; and 0.81 ± 0.43 mm between 1993 and 2003.*

New observations now allow a more quantitative evaluation of changes in the ice sheets of Greenland and Antarctica. The ice sheets in Greenland and Antarctica are likely to be shrinking at present, with thickening in central regions more than offset by increased melting in coastal Greenland, and by acceleration of outlet glaciers of both ice sheets (see Figure TS-16). Assessment of the data and techniques suggests a mass balance of the Greenland Ice Sheet of –50 to –100 Gt (contributing to raising global sea level by 0.21 ± 0.07 mm) per year during 1993–2003, with even larger losses in 2005. The estimated range in mass balance for the Greenland ice sheet over 1961–2003 is between +25 and –60 Gt (0.05 ± 0.12 mm SLE) yr^{-1}. There are greater uncertainties for Antarctica and for older times but assessment of all the data yields an estimate for the overall Antarctic ice-sheet mass balance ranging from +100Gt to –200 Gt (0.14 ± 0.41 mm SLE) yr^{-1} from 1961 to 2003, and from +50 Gt to –200 Gt (0.21 ± 0.35 mm SLE) per year from 1993–2003. Acceleration of mass loss is likely to have occurred in Antarctica, but not so dramatically as in Greenland. Overall, ice sheets have contributed 0.4 ± 0.4 mm yr^{-1} to sea-level rise for the 1993–2003 period and 0.2 ± 0.5 mm yr^{-1} for the 1961–2003 period. [4.6, 4.8]

[INSERT FIGURE TS-16 HERE]

Recent observations show evidence for rapid changes in ice flow in some regions, contributing to sea level rise and suggesting that the dynamics of ice motion may be a key factor in future responses of ice shelves, coastal glaciers, and some ice sheet margins to climate change. Thinning or loss of ice shelves in some near-coastal regions of Greenland and the Antarctic Peninsula, and West Antarctica has been associated with accelerated flow of nearby glaciers and ice streams, suggesting that ice shelves (including short ice shelves of kilometers or tens of kilometers) could play a larger future role in stabilizing or restraining ice motion than previously thought. Both oceanic and atmospheric temperatures appear to contribute to the observed changes. Large summer warming in the Antarctic Peninsula region may have played a role in the subsequent rapid breakup of the Larsen B ice shelf in 2002, by increasing summer meltwater, which drained into crevasses and wedged them open. [4.6]

BOX TS.3.2: ICE SHEET DYNAMICS AND STABILITY

Ice sheets are thick, broad masses of ice formed mainly from compaction of snow. They spread under their own weight, transferring mass towards their margins where it is lost primarily by runoff of surface meltwater or by calving of icebergs into marginal seas or lakes. Ice sheets flow by deformation within the ice or meltwater-lubricated sliding over materials beneath. Rapid basal motion requires that the basal temperature be raised to the melting point by heat from the earth's interior, delivered by meltwater transport, or from the "friction" of ice motion. Sliding velocities under a given gravitational stress can differ by orders of magnitude, depending on the presence or absence of deformable sediment, the roughness of the substrate, and the supply and distribution of water. Basal conditions are well-characterized in few regions, introducing important uncertainties to the understanding of ice sheet stability. [4.6]

Ice flow is often channeled into fast-moving ice streams (which flow between slower-moving ice walls) or outlet glaciers (with rock walls). Enhanced flow in ice streams arises either from higher gravitational stress linked to thicker ice in bedrock troughs, or from increased basal lubrication.
[4.6]

Cold ice discharged across the coast often remains attached to the ice sheet to become a floating ice shelf. An ice shelf moves forward, spreading and thinning under its own weight, and fed by snowfall on its surface and ice input from the ice sheet. Friction at ice-shelf sides and over local shoals slows the flow of the ice shelf and thus the discharge from the ice sheet. An ice shelf loses mass by calving icebergs from the front and by basal melting into the ocean cavity beneath. Available data suggest a 1°C ocean warming could increase ice-shelf basal melt by 10 m per year, but inadequate knowledge of the largely inaccessible ice shelf cavities restricts the accuracy of such estimates. [4.6]

The paleo-record of previous ice ages indicates that ice sheets shrink in response to warming and grow in response to cooling, and that shrinkage can be far faster than growth. Ice-sheets can respond to environmental forcing on very slow time scales, implying that commitments to future changes may result from current warming. For example, a surface warming may take more than 10,000 years to penetrate to the bed and change temperatures there. Ice velocity over most of an ice sheet changes slowly in response to changes in the ice sheet shape or surface temperature, but large velocity changes may occur rapidly on ice streams and outlet glaciers in response to changing basal conditions or changes in the ice shelves into which they flow. [4.6]

Models currently configured for long integrations remain most reliable in their treatment of surface accumulation and ablation, as for the TAR, but do not include full treatments of ice dynamics; thus, analyses of past changes or future projections using such models may underestimate ice-flow contributions to sea level rise, but the magnitude of such an effect is unknown. [8.2]

TS.3.3 CHANGES IN THE OCEAN: INSTRUMENTAL RECORD

The ocean plays an important role in climate and climate change. The ocean is forced by mass, energy and momentum exchanges with the atmosphere. The ocean's heat capacity is about 1000 times larger than that of the atmosphere and the ocean's net heat uptake is therefore many times greater than that of the atmosphere (see Figure TS-17). Global observations of the heat taken up by the ocean are hence a definitive test of changes in the global energy budget. Changes in the amount of energy taken up by the upper layers of the ocean also play a crucial role for climate variations on seasonal to inter-annual time scales, such as El Niño. Changes in the transport of heat and sea-surface temperatures could have important effects upon many regional climates worldwide. Life in the sea is dependent on the biogeochemical status of the ocean and is affected by changes in the physical state and circulation. Changes in ocean biogeochemistry can also feed back into the climate system, e.g., through changes in uptake or release of radiatively active gases such as CO_2. [5.1]

[INSERT FIGURE TS-17 HERE]

Global mean sea level variations are driven in part by changes in density, through thermal expansion or contraction of the ocean's volume. Local changes in sea level also have a density-related component due to temperature and salinity changes. In addition, exchange of water between oceans and other reservoirs (e.g., ice sheets, mountain glaciers, land water reservoirs and atmosphere) can change the ocean's mass and hence contribute to sea level. Sea level change is not geographically uniform because processes such as ocean circulation changes are not uniform across the globe (see Box TS.3.3). [5.5]

Oceanic parameters can be useful for climate change detection, in particular temperature and salinity changes below the surface mixed layer where the variability is smaller and signal-to-noise ratio is higher. Observations analysed since the TAR have provided new evidence for changes in global ocean heat content and salinity, sea level, thermal expansion contributions to sea level rise, water mass evolution and bio-geochemical cycles. [5.5]

TS.3.3.1 Changes in ocean heat content and circulation

The heat content of the world ocean has increased since 1955, accounting for about 90% of the changes in the Earth's energy budget in that time. A total of 7.3 million vertical profiles of ocean temperature allows construction of improved global time series by independent groups of scientists (see Figure TS-18). The data suggest strong interdecadal variations in global ocean heat content superimposed on a multi-decadal trend. Whereas regional warming and cooling could represent redistribution of energy, a sustained global mean ocean warming reflects a change in the Earth's total energy budget. Data coverage limitations require averaging over decades for the deep ocean and observed decadal variability in the global heat content is not fully understood. Analyses of the global oceanic heat budget have been replicated by several independent analysts and are robust to the method used. However, inadequacies in the distribution of data (particularly coverage in the Southern Ocean and South Pacific) could contribute to the apparent decadal variations in heat content. Global ocean heat content is estimated to have increased by 14.1×10^{22} J during the period

1961–2003, equivalent to an average heating rate of 0.2 W m^{-2} (per unit area of the Earth's surface). Estimated warming rates from 1993 to 2003 in the 0–700 m ocean layer range from 0.50 ± 0.16 W m^{-2} to 0.61 ± 0.18 W m^{-2} (per unit area of the Earth's surface). [5.1, 5.2, 5.3]

[INSERT FIGURE TS-18 HERE]

Warming is widespread over the upper 700 meters of the global ocean. While the global trend is one of warming, there are significant regions where the oceans are cooling. Parts of the North Atlantic, North Pacifc and Equatorial Pacific have cooled over the last 50 years (Figure TS-18). The warming is penetrating deeper in the Atlantic ocean basin than in the Pacific, Indian and Southern Oceans, consistent with the deep overturning circulation cell that occurs in the North Atlantic Ocean. The Southern Hemisphere deep overturning circulation shows little evidence of change based on presently available data. However, the upper layers of the Southern Ocean contribute strongly to the overall increase in heat content. The Atlantic has warmed south of 45°N. The changes in the Pacific ocean show ENSO-like spatial patterns linked in part to the Pacific Decadal Oscillation. At least two seas at subtropical latitudes (Mediterranean and Japan/East Sea) are warming. Indirect evidence suggests that the Atlantic meridional overturning circulation has considerable decadal variability but no confirmed trend. [5.2]

TS.3.3.2 Changes in ocean biogeochemistry and salinity

The fraction of CO_2 emissions absorbed by the world's oceans is estimated to have decreased from 42 ± 7% during the period 1750–1994 (see Figure TS-19) to 37 ± 7% during the period 1980–2005. While the fraction of CO_2 emissions taken up by the oceans appears to have decreased, the absolute amount has increased since the 1980s because annual emissions have been increasing. This trend is expected to continue. Observations demonstrate that surface ocean pCO_2 concentrations have increased nearly everywhere roughly following the atmospheric increase, with large regional and temporal variability. [5.4, 7.3]

Estimates of the accumulated ocean storage of anthropogenic CO_2 of 118 ± 19 GtC (circa 1995) (see Figure TS-19) allow corresponding estimates of the net terrestrial release of CO_2 to the atmosphere of 39 ± 38 GtC since 1750. Another estimate of the release of CO_2 of 140 ± 40 GtC due to land use practices (mainly deforestation and agriculture) allows the identification of the former 'missing sink' as a 'residual land sink' with a cumulative storage since 1750 of order 101 GtC and an uncertainty of a factor of 2 (see Table TS-1). [7.3]

[INSERT FIGURE TS-19 HERE]

As a result of this carbon uptake, the surface ocean has become more acidic, with a decrease in pH by 0.1 units since 1750[7]. Decreasing pH increases the volume of ocean water that is undersaturated with respect to the minerals aragonite (a meta-stable form of calcium carbonate) and calcite, which are used by marine organisms to build their shells. Decreasing surface ocean pH and rising surface temperatures both act to reduce the ocean buffer capacity for CO_2 and the rate at which the ocean can take up excess atmospheric CO_2. [5.4, 7.3]

The oxygen concentration of the ventilated thermocline (~100–1000 m) decreased in most ocean basins between 1970 and 1995, consistent with a reduced rate of ventilation linked to upper level warming. Changes in biological activity may also play a role. [5.4]

There is now widespread evidence for changes in ocean salinity in the past half-century. These changes in salinity imply changes in the atmospheric hydrological cycle over the oceans. Ocean waters at high-latitudes (poleward of 50°N and poleward of 70°S) are fresher in the upper 500 m. The sub-tropical latitudes in both hemispheres in the upper 500 m are characterised by an increase in salinity. Regional analyses of salinity display freshening of key high-latitudes water masses such as Labrador Sea Waters (see Figure TS-20), and Antarctic and North Pacific Intermediate Waters, and increased salinity in some of the subtropical gyres such

[7] Acidity is a measure of the concentration of H^+ ions and is reported in pH units, where pH = –log(H^+). A pH decrease of 1 unit means a 10-fold increase in the concentration of H^+, or acidity.

as 24°N. At high-latitudes in the North Hemisphere there is an observed increase in the melt of perennial sea-ice, increased precipitation, and glacial meltwaters, all of which are consistent with fresher high latitude surface waters. Taken together, the observed changes in salinity suggest an intensification in the Earth's hydrological cycle over the last 50 years. [5.3]

[INSERT FIGURE TS-20 HERE]

TS.3.3.3 Changes in sea level

Over the 1961–2003 period, the average rate of global mean sea level rise is estimated from tide gauge data to be 1.8 ± 0.5 mm yr^{-1} (see Figure TS-21). The average thermal expansion contribution to sea level rise for this period is 0.42 ± 0.14 mm yr^{-1}, with significant decadal variations, while the contribution from glaciers, small ice caps, and ice sheets is estimated to represent 0.7 ± 0.5 mm yr^{-1}. The sum of these estimated climate-related contributions of the past 50 years thus amounts to 1.1 ± 0.6 mm yr^{-1}, which is less than the estimated observed value. This discrepancy may be insignificant, given the uncertainties, but it may be explained at least in part by the net contribution of anthropogenic change in terrestrial water storage (principally groundwater mining and impoundment in reservoirs). [5.5]

The longest tide-gauge records available suggest that the rate of sea level rise has accelerated since the 19th century. A recent reconstruction of sea level change back to 1870 using a large set of tide gauges finds a significant acceleration, of 0.013 ± 0.006 mm yr^{-2} over the period 1870–2000, which lies within the range of acceleration of simulated sea level change during this period from climate models. Archaeological data also indicate that sea level rise during the 20th century was larger than the average rise over the centuries of the last millennium. Archaeological and geologic data (such as information from ancient Roman fish ponds) constrain the onset of modern sea level rise to 100 ± 50 years ago and suggest that prior to this onset (i.e. over the previous 2000–3000 years) sea-level rise had an average rate of 0.0–0.2 mm yr^{-1}, i.e., at most six times slower than observed sea-level rise since 1961. [5.5]

[INSERT FIGURE TS-21 HERE]

Precise satellite measurements since 1993 now provide unambiguous evidence of regional variability of sea level change. In some regions, rates are up to several times the global mean rise, while in other regions sea level is falling. The largest sea level rise since 1992 has taken place in the western Pacific and eastern Indian Oceans. (See Figure TS-22). Nearly all of the Atlantic Ocean shows sea level rise during the past decade, while sea level in the eastern Pacific and western Indian oceans has been falling. The pattern of observed sea level change since 1992 is similar to the thermal expansion computed from ocean temperature changes, but different from the thermal expansion pattern of the last 50 years, demonstrating the importance of regional decadal variability. [5.5]

[INSERT FIGURE TS-22 HERE]

The globally averaged rate of sea level rise measured by Topex/Poseidon satellite altimetry during 1993–2003 is 3.1 ± 0.8 mm yr^{-1}. This observed rate for the recent period is close to the total climate-related contributions from thermal expansion and changes in land ice of 2.8 ± 0.8 mm yr^{-1}. The thermal expansion contribution is estimated to represent 1.6 ± 0.6 mm yr^{-1}, while changes in glaciers and ice caps are estimated to have contributed an additional 0.81 ± 0.43 mm yr^{-1} and the Greenland and Antarctic ice sheets are estimated to contribute 0.21 ± 0.07 and 0.21 ± 0.35, respectively (See Table TS-3); the sum of the contributions from all land ice loss is 1.2 ± 0.6 mm yr^{-1}. Climate models generally underestimate the rates of sea level rise during the 1993–2003 period. This may be partly because 1993–2003 was a period of high sea level rise due to internally generated variability, noting that models do not generally simulate as much sea level change due to decadal variability as suggested by thermal expansion data. [5.5, 9.5]

Table TS-3. (Adapted from Table 5.5.2) Contributions to sea level rise for 1993–2003 and for the last 4 decades, compared to observed totals.

	Sea level rise (mm yr^{-1})	
	1961–2003	1993–2003

Thermal expansion	0.42 ± 0.14	1.6 ± 0.6
Glaciers and ice caps	0.51 ± 0.32	0.81 ± 0.43
Greenland ice sheet	0.05 ± 0.12	0.21 ± 0.07
Antarctic ice sheet	0.14 ± 0.41	0.21 ± 0.35
Sum of climate-related terms	1.1 ± 0.6	2.8 ± 0.8
Observed	1.8 ± 0.5 (tide gauges)	3.1 ± 0.8 (satellite altimeter)

BOX TS.3.3: SEA LEVEL

The level of the sea at the shoreline is determined by many factors that operate on a great range of time-scales, from hours (tides and weather) to millions of years (ocean basin changes). The land itself can rise and fall owing to geological processes, including the widespread ongoing response to the melting of ice sheets at the end of the last glacial period, which redistributed the load on the Earth's surface. Land movement has to be allowed for when using coastal measurements of sea level to evaluate the component due to ocean climate change. Since the early 1990s, sea level has also been observed continuously and nearly globally by satellites. Coastal and satellite measurements agree that sea level has been rising during this period on the global average. [5.5]

Thermal expansion of sea water as it warms up is believed to have been the largest contributor to sea-level rise in recent decades. Models are consistent with this and indicate that thermal expansion is expected to contribute most of the sea level rise expected over the next hundred years. Since deep ocean temperatures change only slowly, thermal expansion would continue for many centuries after atmospheric concentrations of greenhouse gases were stabilised. Observations and models show that sea level change has regional variation which is substantial compared with the global average sea level rise, because of the non-uniform distribution of temperature and salinity change in the ocean, changes in winds and in ocean circulation. [5.5]

Global average sea level also rises or falls when water is transferred from land to ocean or vice-versa. Some human activities such as extraction of groundwater and construction of reservoirs contribute to this. However, the major land store is the water frozen in glaciers, ice caps and ice sheets. Sea level was lower during the glacial periods because of the ice sheets covering large parts of the Northern Hemisphere continents. Contraction of glaciers and ice caps is believed to be contributing significantly to current sea level rise and is expected to make a substantial contribution over the next hundred years, but over centuries these sources could be eliminated or restricted to higher, colder altitudes in the mountains. The Greenland ice sheet contains much more ice and could make a large contribution over many centuries. In a warmer climate the Antarctic ice sheet is expected to gain mass on account of increased snowfall, thus lowering sea level, though this could be offset by increased discharge of ice (forming icebergs). In recent years accelerated ice flow and greater discharge has been observed in marginal areas of both ice sheets, tending to increase their sea-level contributions, but the processes responsible are not yet fully understood. [5.5]

Local sea level is affected by climate variability on shorter timescales, for instance associated with El Niño and the North Atlantic Oscillation. This can lead to interannual variations which are much greater than any multidecadal trend. [5.5]

The greatest impacts of sea level change can come through its variability and extremes on the short timescales of days and hours. Tsunamis are large waves caused by submarine earthquakes, landslips and volcanic eruptions - phenomena which are not affected by climate change. But extremes in sea level can be linked to climate in part through tropical cyclones and mid-latitude storms. Low atmospheric pressure and high winds during such storms produce large local sea-level excursions called "storm surges", which are especially serious when they coincide with high tide. Changes in the occurrence of these extreme events are thus affected both by changes in the meteorological phenomena causing them, and in mean sea level. The latter is generally dominant on average, since a particular high-water level at a certain locality may become much more frequent as a result of a modest rise in mean sea level. [5.5]

Observations suggest increases in extreme high water at a broad range of sites worldwide since 1975. Longer records are limited in space, and undersampled in time, so a global analysis over the entire 20th century is not feasible. In many cases, the secular changes in extremes were found to be similar to those in mean sea level, but at others, changes in atmospheric conditions such as storminess were more important in determining long-term trends. Likewise, interannual variability in high water extremes was found to be positively-correlated with regional mean sea level, as well as to indices of regional climate such as ENSO in the Pacific and NAO in the Atlantic. [5.5]

TS.3.4 CONSISTENCY AMONG OBSERVATIONS IN THE INDUSTRIAL ERA

In this section, variability and trends within and across different climate variables including the atmosphere, cryosphere, and oceans are examined for consistency based upon conceptual understanding of physical relationships between the variables. For example, increases in temperature will enhance the moisture-holding capacity of the atmosphere. Changes in temperature and/or precipitation should be consistent with those evident in circulation indices and in glaciers. Consistency between independent observations using different techniques and variables provides a key test of understanding, and hence enhances confidence. [3.9]

Changes in the atmosphere, cryosphere, and ocean strongly support the view that the world is warming. [3.2, 3.9, 4.2, 4.8, 5.2, 5.5]

Both land surface air temperatures and SST show warming. Land regions have warmed at a faster rate than the oceans for both hemispheres in the past few decades, consistent with the much greater thermal inertia of the oceans and differences in land-ocean wetness. [3.2, 3.3]

The warming of the climate is consistent with increases in the number of warm extremes, reductions in the number of daily cold extremes, and reductions in the number of frost days in mid-latitudes. [3.2, 3.8, 3.9]

Surface temperature variability and trends since 1979 are now consistent quantitatively within their respective error bars with those estimated by analyses and uncertainties of satellite retrievals of lower-tropospheric temperatures, provided the latter are adequately adjusted for the known issues of satellite drift, orbit decay, different satellites and stratospheric influence on the tropospheric records. Evidence suggests increasing warming with altitude from 1979 to 2004 from the surface through much of the troposphere in the tropics, cooling in the stratosphere, and a higher tropopause, consistent with expectations from basic physical models and observed increased greenhouse gases and changes in stratospheric ozone. [3.9, 9.4]

Changes in temperature are believed to be consistent with the observed nearly worldwide reduction in mountain glacier mass and extent. Changes in climate consistent with warming are also indicated by decreases in the length of the freeze season of river and lake ice. [4.3, 4.5, 3.9]

Snow cover has decreased in many regions, particularly in the spring season, quantitatively consistent with greater temperature increases in spring as opposed to autumn in mid-latitude regions. These changes are consistent with changes in permafrost, whose temperature has increased by up to 3°C since the 1980s in the Arctic and Subarctic. Snow depth has also decreased in many mountainous regions in a manner consistent with warming. [4.2, 4.7, 3.9]

Observations of sea level rise since 1993 are consistent with observed changes in ocean heat content and the cryosphere. Sea level likely rose by 3.0 ± 0.4 mm/year from 1993–2003, when confidence increases from global altimetry measurements. During this period, glacier, ice cap, and ice sheet melt has increased ocean mass by approximately 1.2 ± 0.6 mm/year, while increases in ocean heat content and associated ocean expansion are estimated to contribute 1.6 ± 0.6 mm/year. This near balance gives increased confidence that the observed sea level rise is a strong indicator of warming, and an integrator of the cumulative energy imbalance at the top of atmosphere. [5.5, 3.9]

Evidence has emerged for consistent changes in circulation, precipitation, and related climate parameters. Consistent changes in climate are reflected in multiple observations of atmospheric circulation patterns, precipitation, salinity, and the hydrologic cycle. [3.5, 3.6, 3.8, 5.2]

Changes in the large-scale atmospheric circulation are apparent. Increasing mid-latitude westerlies have been evident in both hemispheres related to changes in annular modes. In the NH, the NAM and NAO are a major part of the wintertime observed change in storm tracks, precipitation and temperature patterns. In the SH, atmospheric circulation and SAM changes, in association with the ozone hole, have been identified with recent contrasting trends of large warming in the Antarctic Peninsula, and cooling over much of the rest of Antarctica. [3.5, 3.6, 3.8]

Sea-ice extents have decreased in the Arctic, particularly in the summer seasons, and patterns of the changes are broadly consistent with regions showing warming temperatures, although changes in circulation are also a factor. The thickness of Arctic pack ice has also reduced since the 1980s. In contrast to the Arctic, Antarctic sea ice does not exhibit any significant trend since the end of the 1970s, which is broadly consistent with the evolution of surface temperature south of 65°S, which also shows little trend over that period. There has been considerable progress since TAR in understanding how these Arctic and Antarctic trends are linked to changes in both temperature and atmospheric circulation. [4.4, 3.9, 9.5]

Observations are quantitatively consistent with physical understanding regarding the expected linkage between water vapour and temperature, and with acceleration of the hydrologic cycle in a warmer world. Satellite measurements indicate that water vapour has increased over the global oceans by about 1.2% ± 0.3% from 1988 to 2004. Upper tropospheric water vapour has also increased. These data provide important support for the hypothesis of simple physical models that specific humidity increases in a warming world, representing an important positive feedback to climate change. Consistent with rising amounts of water vapour in the atmosphere, there are widespread increases in the numbers of heavy precipitation events and increased risk of flooding events from many land regions, even those where there has been a reduction in total precipitation. Observations of changes in ocean salinity independently support the view that the Earth's hydrologic cycle has changed, in a manner consistent with observations showing greater precipitation and river runoff in high-latitudes, and increased transfer of freshwater from the ocean to the atmosphere at lower latitudes. [3.3, 3.4, 3.9, 5.2]

Although global heavy precipitation has increased, droughts have also increased. While regional droughts have occurred in the past, the widespread spatial extent of current droughts is broadly consistent with an expected acceleration in the hydrologic cycle under global warming. Water vapor increases with increasing global temperature due to increased evaporation over wet regions of the earth’s surface. However, increased continental temperatures are expected to lead to greater evaporation from soils, contributing to drying in dry regions where moisture is limited. Changes in snowpack and snow cover can also reduce available seasonal moisture, and contribute to droughts. [3.3, 3.5, 3.9, 9.5]

Box.TS.3.4 Extreme Weather Events

People affected by an extreme weather event (e.g., the extremely hot summer in Europe in 2003, or the heavy rainfall in Mumbai, India in July 2005) often ask whether human influences on the climate are responsible for the event. A wide range of extreme weather events is expected in most regions even with an unchanging climate, so it is difficult to attribute any individual event to a change in the climate. In most regions, instrumental records of variability typically extend only over about 150 years, so there is limited information to characterize how extreme rare climatic events could be. Further, several factors usually need to combine to produce an extreme event, so linking a particular extreme event to a single, specific cause is problematic.

However, simple statistical reasoning indicates that substantial changes in the frequency of extreme events (and also in the maximum feasible extreme, eg the maximum 24 hour rainfall likely at a specific location) can result from a relatively small shift of the distribution of a weather or climate variable.

Extremes are the infrequent events at the high and low end of the range of values of a particular variable. The probability of occurrence of values in this range is called a probability distribution function (pdf) that is, for many variables, shaped similarly to a “Normal” or “Gaussian” curve (the familiar “bell” curve). TS-Box. 3.4, Figure 1 (taken from Figure 2-32 in the TAR) shows a schematic of a such a pdf and illustrates the effect a small shift (corresponding to a small change in the average or centre of the distribution) can have on the frequency of extremes at either end of the distribution. An increase in the frequency of one extreme (e.g., the number of hot days) will often be accompanied by a decline in the opposite extreme (in this case the number of cold days such as frosts). Changes in the variability or shape of the distribution can complicate this simple picture.

[INSERT FIGURE TS-BOX 3.4, FIGURE 1 HERE]

The SAR noted that data and analyses of extremes related to climate change were sparse. By the time of the TAR, improved monitoring of changes in extremes was available, and climate models were being analysed to provide projections of extremes. Since the TAR, the observational basis of analyses of extremes has increased substantially, so that some extremes have now been examined over most land areas (e.g, daily temperature and rainfall extremes). More models have been used in the simulation and projection of extremes, and multiple integrations of models with different starting conditions (ensembles) now provide more robust information about probability distribution functions and extremes. However, relatively few formal climate change detection and attribution studies focussed on extremes have been completed as yet (Table TS-XXX). For some extremes (eg tropical cyclone intensity), there are still data concerns and/or inadequate models. So some assessments still rely on simple reasoning of how extremes might be expected to change with global warming (eg, warming could be expected to lead to more heat waves).

Table TS-4. Trends, attribution and projections of extreme weather and climate events for which there is evidence of an observed late 20th century trend. Colour coding groups phenomena with similar levels of likelihood of attribution of trend to human influence. Italics indicate cases where no formal detection and attribution study has been completed. [Tables 3.7, 3.8, 9.7, 11.3.3]

Phenomenon	***Likelihood that trend occurred in late 20th century (typically post 1960)***	***Likelihood that observed trend is due to human influence***	***Confidence[a] in trend predicted for 21st century***
Cool days / cool nights / frosts: decrease over mid- and high-latitude land areas	Very likely	Likely	High
Warm days / warm nights: increase over mid- and high-latitude land areas	Very likely	Likely (warm nights)	High
Warm spells / heat waves: increase	Likely	*More likely than not*	High
Proportion of heavy precipitation events: increase over many areas	Likely	*More likely than not*	High (but a few areas with projected decreases in absolute number of heavy events)
Droughts: increase over low-latitudes (and mid-latitudes in summer)	Likely	More likely than not	Moderate – mid-latitude continental interiors in summer (but sensitive to model land-surface formulation)
Tropical cyclones: increase in intensity	More likely than not since 1970	*More likely than not (but with low confidence)*	Moderate (few high-resolution models)
Mid- and high-latitude cyclones: increase in most intense storms; storm tracks move polewards	More likely than not	*Not assessed*	Moderate (intensity not explicitly analysed for all models)
High sea level events: increase (excludes tsunamis)	More likely than not	*Not assessed*	Moderate (most mid-latitude oceans)

Notes:
(a) Confidence terms for projected trends are as follows: "high" means consistency across model projections and/or consistent with theory and/or changes in mean; "moderate" indicates some inconsistencies across model projections or only a few relevant model projections available or analysed.

TS.3.5 A PALEOCLIMATIC PERSPECTIVE

Paleoclimatic observations provide increasing confidence in inferences regarding the past changes in climate on time scales ranging from thousands to millions of years, and, just as important, their uncertainties.

Paleoclimatic studies seldom rely on one method or proxy, but rather several. In this way, results can be cross-verified and uncertainties better understood. Paleoclimatic reconstruction methods make use of measurements of past change derived from ground temperature variations, ocean sediment pore-water change, and glacier extent changes, as well as proxy measurements involving the changes in chemical, physical and biological parameters that reflect past changes in the environment where the proxy grew or existed. It is now well accepted and verified that many biological organisms (e.g., trees, corals, plankton, animals) alter their growth and/or population dynamics in response to changing climate, and that these climate-induced changes are well-recorded in past growth in living and dead (fossil) specimens or assemblages of organisms. Trees, ocean plankton and pollen are some of the best-known and best-developed proxy sources of past climate going back centuries and millennia. Networks of tree-ring width and tree-ring density chronologies are used to infer past temperature changes based on comprehensive calibration with temporally overlapping instrumental data. While these methods are heavily used, remaining research issues concern the distributions of available measurements and how well these sample the globe, and such issues as the degree to which the methods have spatial and seasonal biases. [6.2]

It is very likely that average Northern Hemisphere temperatures during the second half of the 20th century were warmer than any other 50-year period in the last 500 years and likely the warmest in the past 1000 years. The uneven coverage and characteristics of the proxy data mean that these conclusions are most robust over summer, extra-tropical, land areas (see Figure TS-23). These conclusions are based upon proxy data such as the width and density of a tree ring, the isotopic composition of various elements in ice, or the chemical composition of a growth band in corals, requiring analysis to derive temperature information and associated uncertainties. In some cases, temperature and precipitation effects are difficult to separate, or are representative of particular seasons rather than full years, which are among key sources of uncertainty. There are now improved and expanded data since TAR, including e.g., measurements at a larger number of sites, improved analysis of borehole temperature data, as well as more extensive analyses of glaciers, corals, and sediments. However, paleoclimatic data are more limited than the instrumental record since 1850 in both space and time, so that statistical methods are employed to construct global averages, and these are subject to uncertainties. Current data are too limited to allow a similar evaluation to be made for the Southern Hemisphere temperatures prior to the period of instrumental data. [6.6, 6.7]

[INSERT FIGURE TS-23 HERE]

Some post-TAR studies indicate greater multi-centennial Northern Hemisphere variability than was shown in the TAR, due to the particular proxies used, and the specific statistical methods of processing and/or scaling them to represent past temperatures. The additional variability implies cooler temperatures, predominantly during the 12th to 14th, the 17th, and the 19th centuries. These are within the uncertainties quoted in TAR. Some of these changes may be linked to natural forcings due to volcanoes and/or solar activity. For example, reconstructions suggest decreased solar activity and increased volcanic activity in the 17th century as compared to current conditions. One reconstruction suggests slightly warmer conditions than those indicated by the uncertainties that were shown in the TAR, in the 11th century. [6.6]

The ice core CO_2 record over the past millennium provides an additional constraint on natural climate variability. The amplitudes of the preindustrial, decadal-scale Northern Hemisphere temperature changes from the proxy-based reconstructions (<1°C) are broadly consistent with the ice core CO_2 record and understanding of the strength of the carbon cycle-climate feedback. Atmospheric CO_2 and temperature in Antarctica co-varied over the past 650,000 years. Available data suggest that CO_2 acts as an amplifying feedback. This provides evidence for the link between climate and atmospheric CO_2. [6.6]

Paleoclimatic information suggests that many commonly cited past warm periods such as the Medieval Warm Period, Altithermal, and Hypsithermal were distinct only regionally and asynchronously rather than globally. [6.5, 6.6]

Paleoclimate data provides evidence for changes in many regional climates. The strength and frequency of ENSO extremes have varied in association with past changes in orbital forcing, indicating that the impacts of ENSO outside the tropical Pacific may change as background climate and forcings change. There is evidence that the strength of the Asian monsoon, and hence precipitation amount, can change abruptly. The paleoclimate records of northern and eastern Africa and of North America indicate that droughts lasting decades to centuries are a recurrent feature of climate in these regions, so that recent droughts in North America and Northern Africa are not unprecedented. [6.5, 6.6]

Strong evidence from sediment data and from modelling links abrupt climate changes during the last glacial period and glacial-interglacial transition to changes in the Atlantic Ocean circulation. Current understanding suggests that the ocean circulation can become unstable and change rapidly when critical thresholds are crossed. These events have affected temperature by up to 16°C in Greenland and have influenced tropical rainfall patterns. However, they were probably associated with a redistribution of heat between northern and southern hemisphere rather than with large changes in global mean temperature. Ice core data show variations of up to 20 ppm in atmospheric CO_2 concentrations that are broadly in parallel with the temperature evolution in Antarctica. However, such events have not been observed during the past 8000 years. [6.4]

Observations indicate that it is likely that large-scale retreat of the Greenland Ice Sheet and other Arctic ice fields during the previous interglacial (about 125 thousand years ago) contributed between 2.2 and 3.5 meters to a total last interglacial sea level rise of 4 to 6 m above present day. This retreat of the ice on Greenland is believed to be associated with Arctic warming of about 2–4°C which was driven by orbital forcing. See Figure TS-24. Paleoclimate observations also suggest that the Antarctic Ice Sheet likely contributed to the sea level rise of that period. [6.4]

[INSERT FIGURE TS-24 HERE]

TS.4 UNDERSTANDING AND ATTRIBUTING CLIMATE CHANGE

Attribution of climate change is based upon demonstration that the detected change is consistent with model-estimated responses to natural and anthropogenic forcing as well as demonstration that the detected change is not consistent with alternative, physically plausible explanations that exclude these forcings. The first IPCC Assessment Report (IPCC, 1990) contained little observational evidence of a detectable anthropogenic influence on climate. Six years later the IPCC Second Assessment Report (SAR; IPCC, 1996) concluded that the balance of evidence suggested 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 also on its contribution to temperature changes during the latter half of the 20th century. Confidence in the assessment of the human contributions to recent climate change has increased considerably since the TAR, in part because of stronger signals emerging in longer records, as well as an expanded and improved range of observations to more fully address attribution of warming along with other changes in the climate system. Some apparent inconsistencies in the observational record (for example, that for the vertical profile of temperature) have been largely resolved. There have been improvements in the simulation of many aspects of present mean climate and its variability on seasonal to interdecadal timescales, although uncertainties remain [see Box TS.4.1]. Models now employ more detailed representations of processes related to aerosol and other forcings. An anthropogenic signal has now emerged in more aspects of the climate system beyond global mean temperature, including changes in global ocean temperatures and ocean heat content, as well as continental scale temperature trends, temperature extremes, circulation and Arctic sea ice extent. [9.1]

TS.4.1 ADVANCES IN ATTRIBUTION OF CHANGES IN GLOBAL MEAN TEMPERATURE IN THE INSTRUMENTAL PERIOD: ATMOSPHERE, OCEAN AND ICE

Anthropogenic warming of the climate system is widespread and can be detected in temperature observations taken at the surface, in the free atmosphere and in the oceans. [3.2, 9.4]

Evidence of the effect of external influences, both anthropogenic and natural, on the climate system has continued to accumulate since the TAR. Model and data improvements, ensemble simulations, and improved representations of aerosol and greenhouse gas forcing along with other influences lead to greater confidence that most current models reproduce large scale natural internal variability of the atmosphere on decadal and inter-decadal time scales quite well. These advances confirm that past climate variations on large spatial scales have been strongly influenced by external forcings. However, uncertainties still exist in the magnitude and temporal evolution of estimated contributions from individual forcings, due for example to uncertainties in model responses to forcing. Some potentially important forcings such as carbonaceous aerosols have not yet been considered in most formal detection and attribution studies. Uncertainties remain in estimates of internal variability 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 complete descriptions of variability on decadal and longer time scales. See Figure TS-25. [8.3, 8.4, 8.6, 9.2, 9.3, 9.4, Box TS.4.1]

[INSERT FIGURE TS-25 HERE]

It is highly likely (>95%) that the warming observed during the past half century cannot be explained without external forcing. These changes took place at a time when non-anthropogenic forcing factors (i.e., the sum of solar and volcanic forcing) would be expected to have produced cooling, not warming. Attribution studies show that it is very likely that these natural forcing factors alone cannot account for the observed warming [See Figure TS-26, 9.4]. There is also increased confidence that natural internal variability cannot account for the observed changes, due in part to improved studies demonstrating that the warming occurred in both oceans and atmosphere, together with observed ice mass losses. [2.7, 3.2, 5.2, 9.5, 9.7]

[INSERT FIGURE TS-26 HERE]

It is very likely that greenhouse gas forcing has been the dominant cause of the observed global warming over the last 50 years. Without the cooling effect of atmospheric aerosols, it is likely that greenhouse gases alone would have caused more global mean temperature rise than that observed during the last 50 years. A key factor in identifying the aerosol fingerprint, and therefore the amount of cooling counteracting greenhouse warming, is the temperature change through time (see Figure TS-26), as well as the hemispheric warming contrast. Ice core and emission estimates suggest that anthropogenic sulphur emissions have decreased over the past two decades. [6.6, 9.1, 9.2, 9.4]

Widespread warming has been detected in ocean temperatures. Formal attribution studies suggest that it is likely that anthropogenic forcing has contributed to the observed warming of the upper several hundred meters of the global ocean during the latter half of the 20th century. [5.2, 9.5]

The observed ocean warming likely accounts for part of the observed rise in global sea level. There is also evidence of widespread retreat of glaciers, consistent with warming, providing evidence that ice loss has contributed to sea-level rise as well. [5.5]

Anthropogenic forcing has likely contributed to recent decreases in Arctic sea ice extent. Changes in Arctic sea ice are expected given the observed enhanced Arctic warming. Improvements in the modelled representation of sea ice and in ocean heat transport strengthen the confidence in this conclusion. [4.4, 8.3, 9.5]

BOX TS.4.1: EVALUATION OF ATMOSPHERE-OCEAN GENERAL CIRCULATION MODELS

Climate models are the primary tool used for understanding and attribution of past climate variations, and for future projections. Since there are no historical perturbations to radiative forcing that are fully analogous to the perturbations expected over the 21st Century, confidence in the models must be built from a number of indirect methods. In each of these areas there have been substantial advances since the TAR, increasing overall confidence in models. [8.1]

Enhanced scrutiny and analysis of model behaviour has been facilitated by internationally coordinated efforts to collect and disseminate output from model experiments performed under common conditions. This has encouraged a more comprehensive and open evaluation of models, encompassing a diversity of perspectives. [8.1]

Model Formulation. The formulation of climate models has developed through improved spatial resolution, and improvements to numerical schemes and parameterisations (e.g. sea ice, atmospheric boundary layer, ocean mixing). More processes have been included in many models, including a number of key processes important for forcing (e.g., aerosols are now modelled interactively in many models). Most models now maintain a stable climate without use of flux adjustments, although some long-term trends remain in AOGCM control integrations. [8.2, 8.3]

Simulation of present climate There have been improvements in the simulation of many aspects of present mean climate and its modes of variability on seasonal to interdecadal timescales, despite the fact that the artificial constraint of flux adjustment has been eliminated in most models. While significant uncertainties remain in the simulation of cloud and its feedbacks on climate sensitivity, some models have demonstrated improvements in the simulation of certain cloud regimes (notably marine stratocumulus). Simulation of extreme events (especially extreme temperature) has improved, but models generally simulate too little precipitation in the most extreme events.. Simulation of extratropical cyclones has improved, and while tropical cyclones are not resolved, some models can simulate the large scale conditions necessary to infer their frequency and distribution. Improved simulations have been achieved in ocean water mass structure, MOC and ocean heat transport. However most models show some biases in their simulation of the Southern Ocean, leading to some uncertainty in modelled ocean heat uptake when climate changes. [8.3, 8.5, 8.6]

Simulation of modes of climate variability. Models have dominant modes of extratropical climate variability which are like the observed ones, NAM/SAM, PNA, PDO, but they still have problems in representing aspects of them. Some problems remain in the simulation of ENSO (despite an overall improvement), and for other modes of variability, notably the MJO. [8.4]

Simulation of past climate variations. Advances have been made in the simulation of past climate variations. Independently of any attribution of those changes, the ability of climate models to provide a physically self-consistent explanation of observed climate variations on these timescales builds confidence that the models are capturing many key processes for the evolution of 21st Century climate. Recent advances include success in modelling observed changes in a wider range of climate variables over the 20th Century (e.g., regional surface temperatures and extremes, sea ice extent, ocean heat content trends and aspects of the hydrological cycle), and the ability to model the broad features of past, very different climate states such as the Mid-Holocene and Last Glacial Maximum. [6.4, 6.5, 9.3, 9.4, 9.5]

Projections using global climate models. Weather and climate models are used to make predictions and projections of future conditions. Both models use the physical laws governing how the atmosphere, oceans, ice move, change temperature and etc. to make their predictions. Weather forecasters use models to predict the weather for several days or more in advance. These forecasts are highly dependent on the initial conditions at the start of the forecast period. Small errors in the initial conditions or in the model, can lead to bad forecasts. [Question 1.2]

Climate models project the climate for several decades or longer into the future. Since the details of individual weather systems are not being tracked and forecast, the initial conditions are much less important in this case. Expected climate change becomes smaller compared to variability linked to weather at smaller space and time scales. As the area of interest moves from global to regional to local, so the predictability decreases. Uncertainties and errors on the planetary scale generally become magnified and more dominant on the smaller scales, whether these are represented explicitly in the global model or simulated in an embedded regional climate model. In climate projections, the forcings and boundary conditions are very important. These conditions include the amount of sunshine reaching the earth, the amount of particles from volcanic eruptions in the atmosphere, and the amount of anthropogenic gases and particles in the atmosphere. For paleoclimate models, ice sheets are considered boundary conditions in some studies. Small errors in the boundary conditions or in the models, can lead to bad forecasts. [Question 1.2]

A few climate models have been tested for, and shown, skill in initial value prediction, on timescales from weather forecasting (a few days) to seasonal forecasting (annual). The broad predictions of previous climate models, of increasing global temperatures in response to increasing greenhouse gases, have been broadly borne out by observations. This strengthens confidence in near-term climate projections and understanding of related climate change commitments. The fact that, given appropriate initialisation data, these models have hindcast skill also increases confidence that they are adequately representing key processes and teleconnections in the climate system. Tests on the weather forecasting timescale, provide insights into important short timescale processes, such as convection and cloud formation, which can be compared against field observations. These studies increase the understanding of the model skill, and provide insight into biases and drifts when the models are run in climate mode. [8.4]

The possibility of developing model skill measures, based on the above evaluation methods, that can be used to narrow uncertainty by providing quantitative constraints on model climate projections, has been explored for the first time using model ensembles. While these methods show promise, a proven set of measures has yet to be established. [8.1, 9.6, 10.5]

TS.4.2 ATTRIBUTION OF SPATIAL AND TEMPORAL CHANGES IN TEMPERATURE

The observed pattern of tropospheric warming and stratospheric cooling can be attributed to the influence of anthropogenic forcing, particularly that due to greenhouse gas increases and stratospheric ozone depletion. New observations since TAR show that this pattern has also led to an increase in the height of the tropopause that is likely due to greenhouse gas and stratospheric ozone changes. Some uncertainty remains in the estimation of tropospheric temperature trends, particularly in the radiosonde record. [3.2, 3.4, 9.4]

The anthropogenic signal in surface temperature changes has now likely been detected in all inhabited continents. The chance that all regional results in different parts of the globe are spurious and simultaneous is very small, since different regions are affected by different uncertainties in observations, external forcings and internal variability. The ability of models to simulate the temperature evolution on continental scales provides stronger evidence of human influence on the global climate than was available in the TAR. However, difficulties remain in simulating temperature changes in some parts of the world. Changes expected from external forcings become smaller compared to natural internal variability as spatial scales decrease, complicating attribution at sub-continental scales. [7.5, 9.4]

TS.4.3 ATTRIBUTION OF CHANGES IN CIRCULATION, PRECIPITATION AND OTHER CLIMATE VARIABLES

Trends in the Northern and Southern Annular Modes over recent decades, which correspond to sea level pressure reductions over the poles and related changes in atmospheric circulation, are likely related in part to human activity. (See Figure TS-27.) Models reproduce the sign of the Northern Annular Mode trend, but the simulated response is smaller than observed and its structure is somewhat different. 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 that is also consistent with the observed cooling trend in surface climate over parts of Antarctica. These changes in hemispheric circulation and their attribution to human activity imply that anthropogenic effects have likely contributed to changes in mid- and high-latitude patterns of circulation, temperature, and precipitation, as well as changes in winds and storm tracks. However, quantitative effects on surface pressure and precipitation are uncertain because simulated responses to 20th century forcing change agree only qualitatively and not quantitatively with observations of these variables. [9.5, 10.3]

[INSERT FIGURE TS-27 HERE]

Evidence of the impact of external influences on the hydrological cycle is emerging. Observed increases in heavy precipitation appear to be consistent with increases that are expected to occur with warming, although it is unclear whether these changes are distinguishable from natural variability. However, the response to volcanic forcing simulated by some models is detectable in global annual mean land precipitation during the latter half of the 20th century. Observed 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. A number of studies have now demonstrated that changes in land use, due for example to overgrazing and conversion of woodland to agriculture, are not likely to have been the primary cause of Sahelian and Australian droughts. Observed changes in monsoons, storm intensities, and Sahelian rainfall appear to be related to changes in observed sea surface temperatures. While changes in global SSTs are very likely affected by anthropogenic forcing, an association of regional SST changes with forcing has not been established. Changes in rainfall depend not just upon warmer SSTs but also upon changes in the spatial and temporal SST patterns and regional changes in atmospheric circulation, making attribution to human influences complex. [9.5, 10.3]

Surface temperature extremes have likely been affected by anthropogenic forcing. Many indicators of extremes, including the annual numbers of frost days, warm days and cold days, show changes consistent with warming. However, incomplete global data sets for extremes analysis and model uncertainties still restrict the regions and types of detection studies of extremes that can be performed. [9.4]

Anthropogenic forcing has likely contributed to recent decreases in Arctic sea ice extent. Improvements in the modelled representation of sea ice and of ocean heat transport strengthen the confidence in this conclusion. [4.4, 8.3, 9.5]

TS.4.4 PALEOCLIMATE STUDIES OF ATTRIBUTION

Evidence from detection and attribution results indicates that a large fraction of Northern Hemisphere interdecadal temperature variability contained in reconstructions of the past 7 centuries is very likely attributable to natural external forcing. (see Figure TS-28). Such forcing includes: episodic cooling due to known volcanic eruptions, a number of which were larger than those of the 20th century (based on evidence such as ice cores); long-term variations in solar irradiance, such as reduced radiation during the Maunder Minimum; and the emerging greenhouse gas signal. [6.6, 9.3]

[INSERT FIGURE TS-28 HERE]

Attribution studies considering the entire record of the past 700 years support the conclusion that it is likely that greenhouse gas forcing has been the dominant cause of the observed warming of the northern hemisphere over the last 50 years. Insufficient data are available to make a similar southern hemisphere evaluation. [6.6, 9.3]

TS.5 PROJECTIONS OF FUTURE CHANGES IN CLIMATE

Since the TAR, there have been many important advances in understanding climate change projections. An unprecedented effort has been initiated to make available new model results for immediate scrutiny of those outside the modelling centers. A set of coordinated, standard experiments was performed by fourteen Atmosphere-Ocean General Circulation Model (AOGCM) modelling groups from ten countries using 23 models. The resulting model output, analyzed by hundreds of researchers worldwide, forms the basis for much of the current IPCC assessment of model results. For many experiments, ensembles were run to test the sensitivity of the response to initial conditions, and many advances have come from the use of AOGCMs that rely not only on single realizations with single models, but also on multi-member ensembles from single models and from multi-model ensembles. These two different types of ensembles allow new and more robust studies of the range of model results and more quantitative model evaluation against observations, as well as new information on modelled statistical variability. Some studies have also probed probabilistic issues such as the risks of future changes that may be linked to human influences. [8.3, 9.4, 10.1]

A number of methods for providing probabilistic climate change projections, both for global means and geographical depictions have emerged since the TAR. These can be grouped into three broad categories, consisting of: (1) methods based on results of AOGCM ensembles without formal application of observational constraints; (2) methods designed to be constrained by observations of climate change and their uncertainties; (3) methods designed to give results dependent on both observational constraints and distributions of AOGCM results. [10.5]

Advances have occurred through new approaches to combining and comparing models and observations in studies of how climate is likely to change in the near term as well as on longer time scales. In particular, uncertainties have been reduced in estimates of the transient climate response[8], the role of 'commitment' to

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[8] The *Transient climate response* is the change in the global surface temperature, averaged over a 20 year period, centred at the time of CO_2 doubling, i.e., at year 70 in a 1% per year compound CO_2 increase experiment with a global coupled climate model.

near-term climate change, and equilibrium climate sensitivity[9]. The range for equilibrium climate sensitivity was estimated in the TAR (2001) to be between 1.5 to 4.5°C. In previous assessments it was not possible to estimate the probability that climate sensitivity might fall outside that range, nor to determine uncertainties within this range. A growing body of work using new methods, measurements, and models is now available which provides a better quantification of the range as well as the relative probabilities of climate sensitivity. [8.6, 9.6, 10.1]

Model simulations consider the response of the physical climate system to a range of possible future conditions through use of idealised emissions or concentration assumptions. These include "commitment" experiments with greenhouse gases and aerosols held constant at year 2000 levels, CO_2 doubling and quadrupling experiments, SRES marker scenarios for the 2000–2100 period, and three "stabilisation" experiments with greenhouse gases and aerosols held constant after 2100, providing new information on the physical aspects of long term climate change and stabilization. This Working Group I assessment does not consider the plausibility or likelihood of any specific emission scenario and is independent of the IPCC Working Group III assessment of new research on emission scenarios. [10.1, 10.3]

Long-term projections using AOGCMs have been performed for up to two centuries rather than one century as in TAR, allowing improved study of processes that may occur slowly, such as changes in the MOC. Further, a new multi-model dataset using Earth Models of Intermediate Complexity (EMICs) complements AOGCM experiments to extend the time horizon for several more centuries in the future. This provides a greater range of model responses in this assessment as well as new information on climate change commitments over long time scales. Some of these models contain prognostic carbon cycle components, which permit the quantification of likely effects and uncertainties regarding carbon cycle feedbacks. [10.1]

BOX TS.5.1: HIERARCHY OF CLIMATE MODELS

Estimates of change in global mean temperature and sea level rise due to thermal expansion, can be made using simple climate models (SCMs) that represent the ocean-atmosphere system as a set of global or hemispheric boxes, and predict global surface temperature using an energy balance equation, a prescribed value of climate sensitivity and a basic representation of ocean heat uptake. Such models can also be coupled to simplified models of biogeochemical cycles and allow rapid estimation of the climate response to a wide range of emission scenarios. [8.8, 10.5]

Earth System Models of Intermediate Complexity (EMICs) include some dynamics of the atmospheric and oceanic circulations, or parameterisations thereof, and often include representations of biogeochemical cycles, but they commonly have reduced spatial resolution. EMICS can be used to investigate continental scale climate change and long term large scale effects of coupling between Earth system components using large ensembles of model runs or runs over many centuries. For both SCMs and EMICs it is computationally feasible to sample parameter spaces thoroughly taking account of parameter uncertainties derived from tuning to more comprehensive climate models, matching observations, and use of expert judgment. Thus both types of model are well suited to the generation of probabilistic projections for future climate and allow a comparison of the 'response uncertainty' arising from uncertainty in climate model parameters with 'scenario range' arising from the range of emission scenarios being considered. Confidence in the use of EMICs for these purposes has been increased by more in-depth evaluation since the TAR. [8.8, 10.5, 10.7]

The most comprehensive climate models are the AOGCMs. They also include dynamical components describing land suface processes, sea ice, and vegetation. Much progress has been made since the TAR (see Box TS.4.1) and there are over 20 models from different centers available for climate simulations. Most of these models do not require flux adjustments and are thus coupling the various climate system components in a physically consistent way. Although the large scale dynamics of these models are comprehensive, parameterisations are still used to represent unresolved physical processes such as, e.g., the formation of clouds and precipitation, or ocean mixing due to wave processes and the formation of water masses. Uncertainty in parametrisations is one reason why climate projections differ between different AOGCMs. While resolution of AOGCMs is rapidly improving, it is often insufficient to capture fine-scale structure of

[9] *Equilibrium climate sensitivity* refers to the equilibrium change in the annual mean global surface temperature following a doubling of the atmospheric equivalent CO_2 concentration. See the Glossary for more details.

climatic variables in many regions. In such cases the output from AOGCMs can be used to drive limited-area (or regional climate) models that combine comprehensiveness of process representations comparable to AOGCMs with much higher spatial resolution. [8.2]

TS.5.1 Understanding Climate Change Commitment

Model simulations confirm that current greenhouse gas concentrations constitute a commitment (see Box TS.5.2 for a definition) to future climate change. New model results show that if the concentrations of all forcing agents were to be stabilised, further committed changes in atmospheric variables would be expected to follow because of the long response time of the climate system. [10.3, 10.7]

Previous IPCC projections of future climate changes compared to observations now available increase confidence in short term projections and the underlying physical understanding of climate change commitment over a few decades. Projections for 1990-2005 carried out for the IPCC's first and second assessment reports suggested global mean temperature increases of 0.29°C and 0.15°C per decade, respectively[10]. This can now be compared to observed values of about 0.2°C per decade, as shown in Figure TS-29, providing confidence in such short-term projections. Some of this warming was due to commitments to changes linked to the known concentrations of greenhouse gases at the times of those earlier assessments. [1.2, 3.2]

[INSERT FIGURE TS-29 HERE]

If the concentrations of all radiative forcing agents were stabilized today, and assuming no major volcanic eruptions or unusual changes in solar activity, a committed warming of about 0.1°C per decade would be expected in the next several decades. About twice as much warming (0.2°C per decade) would be expected if emissions follow those of any of the SRES marker scenarios. This result is insensitive to the choice among the SRES marker scenarios, none of which consider policy intervention. The range of expected further warming shows limited sensitivity to the choice among SRES scenarios through about 2050 (1.3°C to 1.7°C) with about a quarter of the warming in 2050 being due to the committed climate change expected if all radiative forcing agents were stabilized today (0.4°C). [10.3, 10.5, 10.7]

The commitments to climate change after stabilization of radiative forcing are expected to be about 0.5°C for the first century, and considerably smaller after that. The multi-model average warming when stabilizing concentrations of greenhouse gases and aerosols at year 2000 values after a 20th century climate simulation, and running an additional 100 years, is about 0.5°C at year 2100 (see Figure TS-30), close to the magnitude of warming simulated in the 20th century (i.e., doubling the calculated response in the long term compared to the present, due to the commitment effect). If the B1 or A1B scenarios were to characterize 21st century emissions followed by stabilization at those levels, the additional warming in the 22nd century after commitment is also about 0.5°C. [10.3, 10.7]

[INSERT FIGURE TS-30 HERE]

Box TS.5.2: Climate Change Commitment

Climate change commitment can be defined as the further increase of temperature, or any other quantity in the climate system that continues to change even if the forcing were to be stabilised. Commitment to future climate change is linked to the time scales of adjustment of components of the climate system. The atmosphere adjusts to changes in its boundary conditions on time scales shorter than a month or so. The upper ocean responds with time scales of several years to decades, with the deep ocean and ice sheet response time scales being longer than 1000 years. When the radiative forcing changes, the atmosphere tries to quickly adjust, however because the atmosphere is strongly coupled to the oceanic mixed layer (which in turn is coupled to the deeper oceanic layer), it takes a very long time for the atmospheric variables to come to a statistical equilibrium. During the long periods where the surface climate is changing very slowly, one can consider that the atmosphere is in a quasi-equilibrium state, and most energy is being absorbed by the ocean, so that ocean heat uptake is a key measure of climate change. [10.7]

[10] See IPCC Second Assessment Report, Summary for Policymakers.

An alternative aspect of committed climate change is to identify the effect of past emissions by considering climate model projections in which future emissions are set to zero. This approach considers both (i) the time lags in the responses of the climate system to changes in radiative forcing and (ii) the timescales over which different forcing agents persist in the atmosphere after their emission because of their differing lifetimes. For example, CO_2 is removed from the atmosphere over time scales ranging from decades to millennia, so that concentrations and the resulting forcing decrease slowly in the absence of further emissions implying a long-term commitment to climate change. Typically the climate change commitment due to past emissions includes a period of further increase in temperature for the reasons discussed above followed by a decrease in temperature as the forcing agents are removed from the atmosphere and radiative forcing decreases. [7.3, 10.7]

Both ways of viewing climate change commitment are considered in this report. Where the term *climate change commitment*[11] is used without further qualification it refers to the future commitment under stabilization with radiative forcing held constant. Where *climate change commitment due to past emissions*[12] is used it refers to the commitment in the absence of further emissions. It should be noted that stabilisation of concentrations of radiatively active species is maintained by ongoing emissions that match natural removal rates. The climate change commitment as defined in this manner is linked to the specific stabilization level, which could generally be altered by changing the emission. [Question 10.3]

Due to the long response time scales of the deep ocean and ice sheets, commitment to changes in sea level occurs over a much longer timescale than applies for surface temperature. Sea level rise commitments due to thermal expansion have much longer timescales than warming commitments, owing to slow processes that mix heat into the deep ocean. During 2000–2020, the rate of thermal expansion is insensitive to choice of SRES scenario and is 1.4 ± 0.8 mm yr^{-1}, similar to the observed value during 1993–2003 of 1.6 ± 0.6 mm yr^{-1}. By the end of the 21st century, under the A1B scenario, thermal expansion is projected to contribute 0.23 ± 0.10 m to sea level rise. Excluding possible rapid dynamical changes in the ice sheets, sea level is projected to rise under this scenario by 0.12 ± 0.06 m by 2050 compared to present. By 2100, sea level rise is projected to range from 0.14–0.43 m for the A1B scenario. Continuing thermal expansion would be expected after stabilization of all radiative forcing agents in 2100 at the A1B concentrations, reaching 0.30–0.80 m by 2300. [10.6, 10.7]

Changes in the ice sheets are expected to contribute to future sea level rise commitments (see also Section 5.3 below). If global average temperatures were to be maintained at 3°C above present levels, models suggest that the Greenland ice sheet would contribute to sea level rise by up to 0.4 m per century, or more if ice-flow was accelerated by basal lubrication due to surface meltwater. This level of warming could occur during the 21st century depending upon the climate sensitivity and the emission scenario, and is comparable to that of the last interglacial period about 125,000 years ago, when paleoclimate data suggest that the ice sheet was significantly smaller and and sea level was higher by several meters. The atmosphere is warmer closer to sea level, and it is possible that a critical elevation exists below which the ice sheet can no longer exist. If the Greenland ice sheet were entirely removed, sea level would rise by about 7 m. There is medium likelihood that Greenland deglaciation and the resulting sea level rise would be irreversible even if the global climate was subsequently returned to pre-industrial conditions. [4.1, 6.4, 10.3, 10.7]

The Antarctic ice sheet is projected to behave differently from that of Greenland because it is too cold for widespread surface melting. It is expected to gain ice through increased snowfall in the 21st century, reducing global sea level rise by about 0.1 ± 0.1 m per century. However, in response to weakening of ice shelves by ocean warming or surface melting at the margins, ice flow could accelerate, causing increased discharge which could offset or outweigh the increased snow accumulation (see Section 5.3). A quantitative projection of the consequences is not yet possible, because prognostic models of ice sheets do not include full treatments of basal melting and the dynamics of ice shelves and ice streams, which determine these rapid marginal changes. [4.6, 10.7]

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[11] This term is referred to as *constant concentration commitment* in Section 10.7
[12] This term is referred to as zero emissions commitment in Section 10.7

Commitments to future climate change due to past emissions vary considerably for different forcing agents because of differing lifetimes in the earth's atmosphere (see Box TS.5.2). If the emissions of greenhouse gases were all stopped at the same time, their concentrations in the atmosphere would decrease at very different rates. Some greenhouse gases have relatively short atmospheric lifetimes (decades or less) such as CH_4 and CO, while others such as N_2O have lifetimes of the order of a century, and some have lifetimes of millennia such as SF_6 and PFCs. Atmospheric concentrations of CO_2 do not decay with a single well-defined lifetime if emissions are stopped. While there is rapid removal of some CO_2 over a time scale of a few decades, a fraction of any CO_2 increase persists in the atmosphere for many millennia, representing a very long commitment to climate change due to past emissions. The slow long-term overall buffering of the ocean including $CaCO_3$ sediment feedback require 30,000 – 35,000 years for atmospheric CO_2 concentrations to reach equilibrium. EMICs using coupled carbon cycle components show that the climate change commitment due to past emissions persists over a thousand years, and even over these very long time scales, temperature and sea level do not return to pre-industrial values. [7.3, 10.7]

Model results suggest that 21st century emissions will provide a long-lasting commitment to climate change for multiple centuries, irrespective of later emissions. If emissions were to cease in 2100, emissions that occurred in the 21st century are expected to continue to have an impact even at year 3000, when both surface temperature and sea level rise due to thermal expansion are still projected to be substantially higher than preindustrial. An indication of the long time scales of climate change commitment is obtained by prescribing anthropogenic CO_2 emissions following a path towards stabilization at 750 ppm, but arbitrarily setting emissions to zero at year 2100. In this test case, it takes about 100 to 400 years in the different models for the atmospheric CO_2 concentration to drop from the maximum (ranges between 650 to 700 ppm) to below the level of two times preindustrial CO_2 (~560 ppm), owing to a continuous but slow transfer of carbon from the atmosphere and terrestrial reservoirs to the ocean. (See Figure TS-31.) Atmospheric CO_2 at year 3000 is approximately linearly related to the total amount of carbon emitted in each model, but with a substantial spread among the models. [10.7]

[INSERT FIGURE TS-31 HERE]

TS.5.2 Equilibrium Climate Sensitivity and Transient Climate Response

An expert assessment based on the available constraints from observations and the strength of known feedbacks simulated in the current generation of GCMs indicates that the climate sensitivity is likely to lie in the range 2 to 4.5°C, with a most likely value of about 3°C. Climate sensitivity is very unlikely to be less than 1.5°C. For fundamental physical reasons as well as data limitations, values substantially higher than 4.5°C still cannot be excluded, but agreement with observations and proxy data is generally worse for those high values than for values in the 2–4.5°C range. Because strong constraints with regards to high climate sensitivities (e.g., 95% bound) do not exist, the probability density functions in many approaches have a long tail towards high values exceeding 4.5°C. [9.3, Box 10.2]

Improved understanding has been gained of the possible range in climate sensitivity based upon comparisons of models and several different kinds of observations. Several approaches are used to constrain climate sensitivity during the instrumental period: (i) the transient evolution of surface temperature, upper air temperature, ocean temperature and radiation, of the last 150 years; (ii) the rapid response of the global climate system to changes in the forcing caused by volcanic eruptions; (iii) modelled climatology such as seasonal temperature range. These are complemented by constraints provided by paleoclimate studies such as new reconstructions of the NH temperature record of the past millennium. A few studies based on large ensembles of climate model integrations have shown that testing models' ability to simulate present climate potentially has value in constraining climate sensitivity. Analysis of climate and forcing evolution over the previous centuries and model ensemble studies do not rule out climate sensitivity being as high as 6°C or more. One factor is the extreme range of possible net forcing that could characterize the 20th century if aerosol indirect cooling effects were at the upper end of the uncertainty range, which would cancel out a larger fraction of the positive forcing due to greenhouse gases and require a higher climate sensitivity to explain 20th century temperature trends. (See Figure TS-32.) [8.6, 9.6, Box 10.2]

[INSERT FIGURE TS-32 HERE]

Since the TAR, considerable progress has been made in identifying the key feedbacks that determine equilibrium climate sensitivity. Water vapour provides the largest positive feedback on climate sensitivity, but although there is a spread in model estimates, there is less spread in the combined water vapour + lapse rate feedback, due to an anticorrelation between the two. Recent observational studies support a water vapour + lapse rate feedback close to that expected from models. Recent studies reaffirm that the spread of climate sensitivity among models arises primarily from inter-model differences in cloud feedbacks, with the shortwave response of low- and mid-level clouds identified as a particular source of model differences. Differences between models are now better understood and progress has been made in developing observationally based tests of the components of cloud feedback. Cryospheric feedbacks such as changes in snow cover have been shown to contribute less than cloud or water vapour feedbacks to the overall spread in model estimates of global mean temperature response to forcing, but they can be important for regional climate responses, particularly in high-latitudes. [8.6, 9.4]

The range of transient climate responses among models is smaller than the range in the equilibrium climate sensitivity. The transient climate response is now better constrained by multi-model ensembles and using models of different complexity, its 90% confidence interval is estimated to be 1.2–2.4 °C. The transient response is related to sensitivity in a non-linear way such that high sensitivities do not immediately manifest themselves in the short term response. Transient climate response is strongly affected by the transient ocean heat uptake. Although the ocean models have improved, systematic model biases and the limited availability of ocean temperature data to evaluate transient ocean heat uptake could affect the accuracy of the estimates of transient climate response. [8.6, 9.6, 10.5]

TS.5.3 LARGE SCALE PROJECTIONS FOR THE 21ST CENTURY

Projected 65% probability ranges (mean ±1 standard deviation) for globally-averaged surface warming in 2100 compared to 1980-2000 and including improved carbon cycle feedback uncertainties are estimated to be 1.5–2.8°C, 2.3–4.1°C, 3.0–5.0°C and 3.5–5.8°C for the B1, A1B, A2, and A1FI scenarios, respectively. For many other atmospheric variables such as precipitation or wind changes, the fractions of commitments for those variables would be similar to the commitments for surface temperature. (See Figure TS-33.) [10.5]

[INSERT FIGURE TS-33 HERE]

Many of the patterns of climate change seen in the TAR projections remain in the new generation of models and across ensemble results (see Figure TS-34). Confidence in the robustness of these patterns is increased by the fact that they have not changed while overall model simulations have improved (Box TS.4.1). [8.3, 8.4, 8.5, 10.3]

Projected geographical patterns of warming show largest temperature increases at high northern latitudes and over land, with less warming over the southern oceans and the North Atlantic. In spite of a projected slowdown of the MOC in most models, warming is still expected over the North Atlantic and Europe due to the effects of the increase of greenhouse gases. The projected pattern of zonal mean warming in the atmosphere, displays a maximum in the upper tropical troposphere and cooling in the stratosphere. Zonal mean warming in the ocean is expected to occur first near the surface and in the northern mid-latitudes, with the warming gradually reaching the ocean interior, most evident at high-latitudes where vertical mixing is greatest. It is clear that the projected pattern of change is very similar among the late century cases irrespective of the scenario, with the pattern correlation coefficient for surface air temperature change as high as 0.994 between for example, two different scenarios such as A2 and A1B. As for the zonal means, the fields normalized by the mean warming are very similar for scenarios examined. (See Figure TS-34.) [10.3]

[INSERT FIGURE TS-34 HERE]

Projections of extremes such as changes in the frequency of heat waves are better quantified than in the TAR, due to improved models and a better assessment of model spread based on multi-model ensembles for this probabilistic variable. The TAR concluded there was a risk of increased temperature extremes, with more extreme heat episodes in a future climate. This result has been confirmed and expanded in more recent studies. Future increases in temperature extremes are predicted to follow increases in mean temperature over most of the world except where surface properties change. A multi-model analysis investigated changes in extreme seasonal (DJF and JJA) temperatures (extreme is defined as lying outside the range spanned by 95% of the cases) based on simulations of 14 models for 3 scenarios. By the end of the 21st century; the projected probability of extreme warm seasons occurring rises above 90% in many tropical areas, and reaches around 40% elsewhere. Several recent studies have addressed possible future changes in heat waves, and found that in a future climate heat waves are expected to be more intense, longer-lasting and more frequent. Based on an 8 member multi-model ensemble, heat waves are simulated to have been increasing for the latter part of the 20th century, and are projected to increase globally and over most regions. [10.3]

Projections generally show increases in the tropical precipitation maxima (such as the monsoon regimes) and over the tropical Pacific in particular, with general decreases in the subtropics and some mid-latitude areas, and increases at high-latitudes. Globally averaged mean specific humidity, evaporation and precipitation are projected to increase. The intensity of rainfall events is also projected to increase, and precipitation extremes are predicted to increase more than does the mean in many areas, particularly outside the tropics. As a consequence of warmer temperatures and increased water vapor, increased moisture is expected to be transported to the intertropical convergence zones, increasing precipitation there. Projections also suggest a tendency for summer drying of the mid-continental areas during summer, albeit with some variation in detail across models. However, some general principles can be identified. In particular, models suggest a future poleward expansion of the subtropical highs, and poleward displacement of the mid-latitude westerlies along with the associated storm tracks. These changes result in drying tendencies associated with the land areas downstream of these centers. [See Section 5.5] [10.3, 11.1, 11.3]

For a future warmer climate, models project a decline in frequency of cold air outbreaks relative to the present of 50 to 100% in NH winter in most areas. Results from a 9 member multi-model ensemble show simulated decreases in frost days for the 20th century continuing into the 21st century globally and in most regions. A quantity related to frost days is growing season length and this has been projected to increase in future climate. [10.3, Question 10.1]

Models show a weakening of the Atlantic MOC over the 21st Century, but there is a wide range in the model estimates of the magnitude of the weakening, ranging from small values up to 60% using some SRES emission scenarios. The reduction of the Atlantic MOC is due to the combined effects of an increase of high latitude temperatures and high latitude precipitation, both of which reduce the density of the surface waters in the North Atlantic. No models run for this assessment suggest an abrupt MOC shutdown during the 21st Century, but some models of reduced complexity suggest MOC shutdown as a possible long-term response to sufficiently strong warming. Very few AOGCM studies have included the impact of additional fresh water from melting of the Greenland Ice Sheet, but those that have do not suggest that this will lead to a complete MOC shutdown. Taken together, it is likely that the MOC will reduce, perhaps associated with a significant reduction in Labrador Sea Water formation, but very unlikely that the MOC will undergo an abrupt collapse during the course of the 21st century. Temperatures over the North Atlantic Ocean and Europe are projected to warm despite such MOC changes, due to the much larger radiative effects of the increase of greenhouse gases. No models suggest an abrupt MOC shutdown during the 21st Century. The likelihood of longer-term changes cannot be evaluated with confidence. The few available simulations with models of different complexity rather suggest a centennial scale slow-down. Recovery of the MOC is likely if the radiative forcing is stabilised but would take several centuries. Systematic model comparison studies have helped establish some key processes that are responsible for variations between models in the response of the ocean to climate change (especially ocean heat uptake and MOC changes). [8.7, 10.3]

Changes are expected to occur in the future intensities of extratropical storms, with associated changes in wave heights. More evidence has been presented since the TAR to support the suggestion that the most intense extratropical storms will become more frequent (though with fewer storms overall). As a consequence of the increase in intensity of storms, extreme wave heights are likely to increase over most of the mid-latitude oceans. [10.3]

Recent studies with improved global models do not suggest significant increases in the total number of future tropical cyclones (typhoons and hurricanes). However, models suggest increased storm intensities, with larger peak wind speeds and more intense precipitation. One simulation with very high resolution and able to represent strong winds in the central part of tropical cyclones shows that the number of strongest tropical cyclones (maximum wind speed >45 m s^{-1}) increases though the number of weaker ones decreases. The observed 20th century increase in the proportion of very intense hurricanes is in the same direction but much larger than simulations by theoretical models. [10.3]

Sea level pressure is projected to increase over the subtropics and mid-latitudes, and decrease over high-latitudes associated with an expansion of the Hadley Circulation and a poleward shift of the storm tracks and annular mode changes (NAM/NAO, and SAM). There is a positive trend of the NAM/NAO as well as the SAM index projected by many models. The magnitude of the projected increase is generally greater for the SAM, and there is considerable spread among the models. [10.3]

A majority of models show a mean El Niño-like response pattern in the tropical Pacific, with the central and eastern equatorial Pacific sea surface temperatures warming more than the western equatorial Pacific, with a corresponding mean eastward shift of precipitation. Future changes of ENSO interannual variability differ from model to model. Some models that show increases more successfully simulate present day characteristics of ENSO, though the large inter-model differences in future changes of El Niño amplitude, and the inherent century-timescale variability of El Niño in the models, preclude a definitive assessment. [10.3]

Under several different scenarios (SRES A1B, A2 and B1) large parts of the Arctic ocean are expected to no longer have year-round ice cover by the end of the 21st century. Arctic sea ice is responding sensitively to warming. While changes in winter sea ice cover are moderate, late summer sea ice is projected to disappear almost completely towards the end of the 21st century. The reduction is accelerated in the Arctic where a number of positive feedbacks in the climate system accelerate the melting of sea ice. The ice albedo feedback allows open water to receive more heat from the sun during summer, and the increase of ocean heat transport to the Arctic through the advection of warmer waters and stronger circulation further reduces ice cover. Model simulations indicate that the September sea ice cover reduces substantially and generally evolves on the same time scale as global warming. Some models project sea ice cover to disappear entirely in the high forcing A2 scenario in the latter part of the 21st century. Sea ice is also projected to reduce in the 21st century in the Antarctic. (See Figure TS-35.) [10.3, Box 10.1]

[INSERT FIGURE TS-35 HERE]

As the climate warms, it is expected that land ice will also decrease. Glaciers and ice caps are projected to lose mass (indicating a dominance of melting over precipitation increases) and contribute to sea level rise as documented for the previous generation of models in the TAR. Snow cover is also projected to decrease. Thawing of the upper layer of permafrost is projected to be as much as 90% in the A2 scenario. [10.3]

The Greenland ice sheet is projected to lose mass in the 21st century because increased melting will exceed increased snowfall. The observation in west-central Greenland of seasonal variation in ice flow rate and of a correlation with summer temperature variation has prompted speculation that surface meltwater is able to reach the bed and lubricate the ice flow; this would increase its sea level contribution, but other explanations for flow variability cannot be discounted.

The Antarctic ice sheet will gain mass, because of increased snowfall without substantial surface melting, unless there are larger accelerations in ice flow of the kind that appear to be presently taking place in some West Antarctic ice streams. These accelerations are likely to have been caused by enhanced basal melting of ice shelves resulting from ocean warming, and could occur in other areas as well, but the recent dynamical imbalance would have to increase by a factor of about five by 2100 to outweigh the projected increased snow accumulation in scenario A1B, for example.

Global average sea level is projected to rise under the A1B scenario by 0.12 ± 0.06 m by 2050 compared to present, excluding possible rapid dynamical changes in the ice sheets. By 2100, sea level rise of 0.29 ± 0.15 m is projected for the A1B scenario,

Models indicate that sea level rise during the 21st century will not be geographically uniform. AOGCMs give an average spatial standard deviation of 140 mm, which is substantial compared with the global average sea level rise. The geographical patterns of sea level change arise mainly from changes in the 3-D distribution of heat and salinity in the ocean and consequent changes in ocean circulation. [10.6]

TS.5.4 Coupling Between Climate Change and Changes in Biogeochemical Cycles

Models that treat the coupling of the carbon cycle to climate change indicate that this leads to a positive feedback. Since the TAR several new fully coupled carbon cycle-climate model based projections have been performed and inter-compared. The coupled versions of these simulations all exhibit higher atmospheric CO_2 increases and stronger climate change than in the uncoupled cases. This is because, in the coupled cases, climate change in general reduces land and ocean carbon uptake. All models show positive climate-carbon cycle feedbacks but their strength varies markedly. For the SRES A2 scenario, the increase in atmospheric CO_2 concentration over the 21st century is from 4% to 44% higher, depending on the model used, when feedbacks are included. This positive feedback effect could lead to as much as 1.2°C of added warming by 2100 for higher SRES emission scenarios. Alternatively it reduces the total emissions consistent with a given CO_2 stabilization level, although there are still uncertainties due for example to limitations in the understanding of the dynamics of land ecosystems and soils [7.3, 10.4]

Increasing atmospheric CO_2 concentrations lead directly to increasing acidification of the surface ocean. Projections based on SRES scenarios give reductions in pH of between 0.14 and 0.35 units in the 21st century (depending on scenario), adding to the present decrease of 0.1 units from pre-industrial times. Ocean acidification will eventually lead to undersaturation and dissolution of calcium carbonate in parts of the surface ocean. While Southern Ocean surface waters are projected to first exhibit undersaturation with regard to $CaCO_3$, low latitude regions will be affected as well. These changes will not only influence the global carbon cycle, but also threaten marine organisms that form their exoskeletons out of $CaCO_3$, which are essential components of the marine food web. [Box 7.3, 10.4]

Future concentrations of many greenhouse gases and their precursors are expected to be coupled to future climate change. Insufficient understanding of the causes of recent variations in the CH_4 growth rate suggest very large uncertainties in future projections for this gas in particular. Since biogenic CH_4 production and emission from major sources (wetland, landfill, rice agriculture) are temperature-dependent, climate change scenarios with a warmer atmosphere suggest enhanced emissions from these sources. Wetlands and rice agriculture are identified as the CH_4 sources that are most sensitive to climate change and thus will likely contribute most to changes in global CH_4 budgets in the future. Changes in temperature, humidity, and clouds could also affect biogenic emissions of other ozone precursors such as volatile organic compounds. Climate change is also expected to affect tropospheric ozone through changes in chemistry and transport. Climate change could induce changes in OH through changes in humidity, and could also alter stratospheric ozone and hence solar ultraviolet in the troposphere. [7.4]

Future emissions of many aerosols and their precursors are expected to be coupled to climate change. Estimates of future changes in dust emissions under several climate and land-use scenarios suggest that dust emissions in the next century are uncertain. Estimate of future changes in dust emissions under several climate and land use scenarios suggest that the effects of climate change are more important in controlling future dust emissions than changes in land-use. The biogenic emission of volatile organic compounds, a significant source of secondary organic aerosols, is known to be highly sensitive to (and increase with) temperature. However, aerosol yields decrease with temperature and the effects of changing precipitation and physiological adaptation are uncertain. Thus change in biogenic secondary organic aerosol production in a warmer climate could be considerably lower than the response of biogenic volatile organic carbon emissions. Climate change may affect fluxes from the ocean of dimethylsulfide , which is a precursor for some sulfate aerosols, and sea salt aerosols, however, the effect on temperature and precipitation remain very uncertain. [7.5]

Studies suggest that increases in aerosols between pre-industrial times and the present would have tended to decrease globally-averaged evaporation and precipitation while other influences such as greenhouse gas increases exerted the opposite effects. Projections indicate that both may increase with climate change over the next century if increased global mean warming due to black carbon and greenhouse gases dominates over the sulphate cooling. [7.5]

While the warming effect of CO_2 represents a commitment over many centuries, aerosols have time scales of only a few days, so that the negative radiative forcing due to aerosols is expected to change rapidly in response to any changes in emissions of aerosols or aerosol precursors. Climate change in response to the total net forcing of greenhouse gas plus aerosols therefore depends on the intensity and stability of aerosol cooling. Because aerosols are believed to be exerting a substantial negative radiative forcing at present, the total future net global mean forcing could be subject to rapid changes if aerosol emissions were to be altered. The hypothetical removal from the atmosphere of the entire current burden of anthropogenic sulphate aerosol particles would produce a rapid increase of about 0.8°C within a decade or two in the globally averaged temperature. Changes in aerosols would also be expected to influence precipitation. Thus, environmental strategies aimed at stabilization or climate change commitment below a prescribed threshold would require consideration not only of greenhouse gas emissions, particularly of CO_2, but also of measures that may be implemented to improve air quality. [Box 7.4, 7.6, 10.7]

Future climate change could degrade air quality. Climate change affects air quality by modifying the dispersion rate of pollutants, the chemical environment for ozone and aerosol generation, and the strength of emissions from the biosphere, fires, and dust, although these effects are highly uncertain and likely to vary from one region to another. [Box 7.4]

TS.5.5 REGIONAL SCALE PROJECTIONS

A direct consequence of improved GCM performance has been the improved simulation of regional climates using regional climate models (RCMs) nested in GCMs, and from empirical downscaling techniques forced by GCMs. These approaches show improving skill in deriving accurate local-scale present day climate representations from the GCM-scale atmospheric forcing. For some of the regions of the world it is now possible to make robust statements about projected changes, either based directly on AOGCM simulations and/or from downscaling methods. The strength and specificity of these statements is region dependent. This represents a significant advance over the TAR. Regional projections of temperature and precipitation are comparable in magnitude to the TAR, but with greater confidence in some regions. [11.1]

For each of the continental regions the projected warming over 2000-2050 based upon SRES scenarios is greater than the global average rate, and greater than the observed rate of warming over the past century. The warming projected in the next few decades of the 21st century would substantially exceed estimated 20th century natural forced and unforced temperature variability over all inhabited continents (Figure TS-36). As was shown in Figure TS-30, the simulated warming over this period is not very sensitive to the choice of scenarios across the SRES set, which do not involve climate policy options. On longer time scales the choice of scenario is more important as shown in Figure TS-34. [9.3, 11.1]

[INSERT FIGURE TS-36 HERE]

There remain a number of major sources of uncertainty limiting our ability to project regional climate change. At the regional scale some uncertainties from the projections presented in this report relate to positional boundaries between climate regimes; within core climate regions, models have greater coherence, but there is uncertainty in the spatial positioning of regime boundaries. The consistency in simulated precipitation change across GCMs varies considerably with region and season. There are some important climate processes which have a significant effect on regional climate, but for which the climate change response is still poorly known. These include ENSO, the NAO, blocking, the thermohaline circulation, and changes to global tropical cyclone distribution. For those regions which have strong topographical controls on their climatic patterns there is often insufficient climate change information at fine spatial resolution. In some regions there has been only very limited study of key aspects of regional climate change, particularly with regard to extreme events. Further, the expected signal of forced climate change becomes smaller compared to internal variability at smaller space and time scales. [Box 11.1, 11.3]

Other regional climate changes that are expected include: [11.3]

- *Africa:* decreases in annual rainfall in portions of North Africa and Northern Sahara and in winter rainfall for regions of south western Africa.

- *Mediterranean and Europe:* winter minimum temperatures increase by more than winter average temperatures in northern Europe; annual precipitation increases in most of northern Europe and decreases in most of the Mediterranean area; extremes of daily precipitation increase in northern Europe; the annual number of precipitation days decreases in the Mediterranean area; snow season and depth decreases.

- *Asia:* warming well above the global mean in Central Asia, the Tibetan Plateau and Northern Asia, above the global mean in Eastern Asia and South Asia, and similar to the global mean in Southeast Asia; summer heat waves / hot spells of longer duration, more intense, and more frequent in Eastern Asia; fewer very cold days in East and South Asia; winter precipitation increases in Northern Asia, East Asia and the Tibetan Plateau; increases in the return frequency of intense precipitation events in parts of South Asia, East Asia, and Southeast Asia.

- *North America:* increases in lowest winter temperatures greater than in average winter temperature in northern North America; increases in annual precipitation in northern parts of North America with decreases in the length of the snow season and snow depth.

- *Central and South America:* annual precipitation decreases along the southern Andes and summer precipitation increases in south eastern South America; all regions warm by more than the global mean, except in southern South America where warming is projected to be similar to the global mean.

- *Australia - New Zealand:* an increase in rainfall in the west of the South Island of New Zealand; an increase in frequency of extreme high daily temperatures, a decrease in the frequency of cold extremes, and an increase in the frequency of extreme precipitation; increased risk of drought in southern areas of Australia.

- *Polar:* Arctic warming greater than global mean warming; annual Arctic precipitation increases; temperature increases more slowly in the Antarctic than in the Arctic.

- *Small Islands:* islands in regions of enhanced sea level rise to be vulnerable to coastal erosion and flooding.

There is now greater confidence in projected patterns of expected future changes in precipitation than in the TAR. Observed patterns of changes in land precipitation with latitude appear to be qualitatively consistent with simulations of the 20th century, suggesting a human contribution to the observed changes. Decreases in precipitation are robustly predicted by more than 90% of the simulations by the end of the 21st century for the northern and southern subtropics. Decreases are also expected for parts of western North and South America, and southern Europe, with increases expected at higher latitudes poleward of about 55° latitude. Even in those regions where precipitation increases, the land may become drier due to increases in evaporation. These changes are in accord with changes in physical processes related to increased atmospheric moisture availability contributing to precipitation in the inter-tropical convergence zones, and a poleward expansion of the subtropical highs influencing the extra-tropics and mid-latitudes. (See Figure TS-37). [8.3, 10.3, 11.3]

[INSERT FIGURE TS-37 HERE]

TS.6 ROBUST FINDINGS AND KEY UNCERTAINTIES

TS.6.1 CHANGES IN HUMAN AND NATURAL DRIVERS OF CLIMATE

Robust Findings:

- It is virtually certain that present levels of LLGHGs and their associated positive radiative forcing (warming effect) are far above the range of natural variability over the last 650,000 years.

- It is virtually certain that human activities have been the dominant cause of increases in greenhouse gases and aerosols in the atmosphere over the last 250 years.

- The average decadal scale rate of increase in radiative forcing from CO_2, methane and nitrous oxide is larger than at any time during the past 20,000 years.

- It is very likely that a significant fraction (42 ± 7%) of all the CO_2 released into the atmosphere has been absorbed by the oceans. Acidity of the ocean has increased in surface waters.

- Anthropogenic aerosols produce a net negative radiative forcing (cooling effect) with a greater magnitude in the northern hemisphere than in the southern hemisphere.

- Solar and land cover contributions to radiative forcing are small relative to the contribution of greenhouse gases over the industrial period.

- The net effect of human activities in the last 250 years has very likely exerted a warming influence on climate.

- Global Warming Potentials provide a robust metric for comparing the potential climatic effect of emissions of different LLGHGs.

***Key Uncertainties*:**

- The full range of processes leading to modification of cloud properties by aerosols is not well understood and the magnitudes of associated indirect radiative effects are poorly determined.

- The radiative forcing due to stratospheric water vapour and the net radiative forcing due to surface property changes are not well quantified.

- The contribution of past solar changes to radiative forcing on the time scale of centuries has not been measured directly.

TS.6.2 OBSERVATIONS OF CHANGES IN CLIMATE

TS.6.2.1 Atmosphere and Surface

Robust Findings:

- Global mean temperatures continue to rise. Ten of the eleven warmest years since 1850 have occurred since 1995.

- Rates of surface warming increased in the mid-1970s and the land surface has been warming at about double the rate of the ocean surface since then.

- Changes in surface temperature extremes are consistent with warming of the climate.

- Precipitation over land increased north of 30°N over the period 1901-2005 and decreased in the tropics and subtropics since the 1970s.

- Substantial increases have occurred in the number of heavy precipitation events.

Key Uncertainties:

- Radiosonde records are much less complete spatially than surface records and evidence suggests a number of radiosonde records are unreliable, especially in the tropics.

- While changes in large-scale atmospheric circulation are apparent the quality of analyses is best only after 1979, making separation of forced change and natural variability difficult.

- Surface and satellite observations disagree on total and low-level cloud changes over the ocean.

TS.6.2.2 Snow, Ice and Frozen Ground

Robust Findings:

- The amount of ice on the Earth is decreasing. There has been widespread retreat of mountain glaciers since the end of the 19th century. The rate of mass loss from glaciers and the Greenland ice sheet is increasing.

- The extent of spring time snow cover has declined. Seasonal river and lake ice duration has decreased the past 150 years.

- Since 1978, annual mean Arctic sea ice extent has been declining and summer minimum Arctic ice extent has decreased more rapidly.

- Ice thinning occurred in the Antarctic Peninsula and Amundsen shelf ice during the 1990s. Tributary glaciers have accelerated and complete break-up of Larsen-B ice shelf occurred in 2002.

- Permafrost regions warmed by up to 3°C since the 1980s. Seasonal frozen ground has been thinning and decreasing in area.

Key Uncertainties:

- There is no compilation of in-situ snow data prior to 1960. Well-calibrated snow water equivalent data are not available for the satellite era.

- There are no global sea ice thickness observations.

- Uncertainties in glacier mass loss estimates arise from limited global inventory data, incomplete area-volume relationships, and imbalance in geographic coverage.

- Mass balance estimates for ice shelves and ice sheets, especially for Antarctica, are limited by calibration and validation of changes detected by satellite altimetry and gravity measurements.

TS.6.2.3 Oceans and Sea Level

Robust Findings:

- It is virtually certain that global temperature (or heat content) of the oceans has increased since 1955.

- It is very likely that large-scale regionally coherent trends of salinity have been observed over recent decades: freshening in polar regions, increased salinity in the mid-latitudes and freshening on the equator.

- It is virtually certain that global average sea level rose during the 20th century. It is very likely that the rate of rise in the 19th century was smaller than the 20th century. During 1993–2003 sea-level rose more rapidly than during 1961–2003. Thermal expansion of the ocean and loss of mass from glaciers and ice caps made substantial contributions to the 1993–2003 rate of rise.

- The rate of sea level change over recent decades has not been geographically uniform.

Key Uncertainties:

- Limitations in ocean sampling (particularly in the Southern Hemisphere) mean that decadal variability in global heat content, salinity, and sea-level changes can only be evaluated with moderate confidence.

- There is low confidence in observations of trends in the meridional overturning circulation including the Atlantic.

- Global average sea level rise for the last 50 years is significantly larger than can be explained by thermal expansion and land ice melting.

TS.6.2.4 Paleoclimate

Robust Findings:

- During the last interglacial the Arctic was 3 to 4°C warmer than at present and global sea level was likely more than 4 m higher due primarily to ice sheet retreat.

- A number of past abrupt climate changes were very likely linked to changes in Atlantic Ocean circulation and had global implications.

- There is no evidence for a natural interglacial climate cycle that could explain recent global warming, or that the current warming will be mitigated by a natural cooling trend.

- Biogeochemical and biogeophysical feedbacks have amplified climatic changes in the past.

- It is very likely that average Northern Hemisphere temperatures during the second half of the 20th century were warmer than in any other 50-year period in the last 500 years and likely that this was also the warmest 50-year period in the past 1000 years.

- Droughts lasting decades to centuries are a recurrent feature of climate in North America and northern Africa under a wide range of climate forcing.

Key Uncertainties:

- A comprehensive mechanistic explanation of observed glacial-interglacial variations in climate and greenhouse gases remains to be articulated.

- Mechanisms of past abrupt climate change or key climate thresholds are not well understood. This limits confidence in the ability of climate models to simulate realistic abrupt change.

- The processes by which ice sheets disintegrated in the past, and the rates of such change, are not well known.

- Knowledge of climate variability over the last 1000 years in the Southern Hemisphere and tropics is severely limited by the lack of paleoclimatic records.

- Available millennial-length northern hemisphere temperature reconstructions have different amplitudes of temperature change and the relationship between these differences and choices of proxy data or statistical calibration methods, has not yet been determined.

- Lack of data for temperature proxies in the last 20 years limits understanding of how these respond to rapid global warming and restricts determination of role of other environmental changes.

TS.6.3 UNDERSTANDING AND ATTRIBUTING CLIMATE CHANGE

Robust Findings:

- It is highly likely (>95%) that observed warming over the past 50 years cannot be explained without external radiative forcing. It is very likely that the warming was not due to known natural causes and that greenhouse gas forcing has been the dominant cause.

- It is likely that greenhouse gases would have caused more warming than observed over the last 50 years with some warming offset by net cooling caused by natural and other anthropogenic factors.

- It is likely that anthropogenic forcing has warmed the upper ocean during the last 50 years.

- Natural external forcings (volcanic and solar effects) explain a substantial fraction of inter-decadal variability in northern hemisphere pre-industrial temperature reconstructions over the last 700 years.

Key Uncertainties:
- Confidence in attributing some climate change phenomena to anthropogenic influences is currently limited by uncertainties in radiative forcing and in observations.

- There is less confidence in understanding of forced changes in precipitation and surface pressure than there is of temperature.

- The range of attribution statements is limited by the absence of formal detection and attribution studies, or their very limited number, for some phenomena (e.g., some types of extreme events).

TS.6.4 Projections of Future Changes in Climate

TS.6.4.1 Model Evaluation

Robust Findings:
- There is considerable confidence that models provide useful projections of future climate change, particularly at global scales, due to their foundation on accepted physical principles and their ability to reproduce observed features of recent climate and past climate changes during periods of very different forcings.

- Confidence in models has increased due to:
    - improvements in the simulation of many aspects of present climate including important modes of climate variability, and extreme hot and cold spells;
    - improved model resolution, numerics and parameterisations, and inclusion of additional processes;
    - more comprehensive diagnostic tests, including tests of model ability to forecast on time scales from days to a year, when initialized with observed conditions;
    - enhanced scrutiny of models and expanded diagnostic analysis of model behavior facilitated by internationally coordinated efforts to collect and disseminate output from model experiments performed under common conditions.

Key Uncertainties:
- A proven set of model metrics comparing simulations with observations, that might be used to narrow the range of plausible climate projections, has yet to be developed.

- Climate drift, primarily in the deep ocean, remains an issue in most AOGCM control simulations.

- Models differ considerably in their estimates of the strength of different feedbacks in the climate system.

- Problems remain in the simulation of some modes of variability, notably the MJO and extreme precipitation.

- Systematic biases have been found in most models' simulations of the Southern Ocean which is linked to uncertainty in transient climate response.

TS.6.4.2 ***Sea Level***

Robust Findings:

- Sea level will continue to rise in the 21st century because of thermal expansion and loss of land ice. Sea level rise was not geographically uniform in the past and will not be in the future.

- Sea level rise due to thermal expansion and loss of mass from ice sheets would continue for centuries or millennia after stabilisation of climate. Projected warming due to emission of greenhouse gases during the 21st century according to some SRES scenarios is likely to be sufficient to eliminate the Greenland ice sheet over the following 1000 years or more.

Key Uncertainties:

- Models do not yet exist that address all processes that may contribute to large rapid dynamical changes in the West Antarctic and Greenland ice sheets that could increase the discharge of ice into the ocean.

TS.6.4.3 *Equilibrium Climate Sensitivity*

Robust Findings:

- Equilibrium climate sensitivity is likely to be in the range 2–4.5°C with a most likely value of about 3°C, based upon multiple observational and modelling constraints.

- There is a good understanding of the origin of differences in equilibrium climate sensitivity found in different models.

- New observational and modelling evidence strongly supports a combined water vapour-lapse rate feedback of a strength comparable to that found in GCMs.

Key Uncertainties:

- Cloud feedbacks are a primary source of inter-model differences in equilibrium climate sensitivity, with low cloud being the largest contributor.

- The magnitude of long-term cryospheric feedbacks, such as partial or complete melting of the Greenland Ice Sheet, remains uncertain, contributing to the range of model climate responses.

TS.6.4.4 *Global Projections*

Robust Findings:

- All models project temperature increases in a narrow range of 0.64–0.70°C, averaged over 2011–2030 relative to 1980–1999, regardless of emission scenario.

- Geographical patterns of projected warming show greatest temperature increases at high northern latitudes and over land, with less warming over the southern oceans and North Atlantic.

- Changes in precipitation now show robust large-scale patterns: precipitation generally increases in the tropical precipitation maxima, decreases in the subtropics, and increases at high latitudes as a consequence of a general intensification of the global hydrological cycle.

- As the climate warms, snow cover and sea ice extent decrease; glaciers and ice caps lose mass and contribute to sea level rise. Sea ice reduces in the 21st century both in the Arctic and Antarctic. The reduction is accelerated in the Arctic where permafrost is substantially reduced in the upper layers of the soil.

- The meridional overturning circulation in the Atlantic slows down in response to warming and freshening of the upper layers of the North Atlantic Ocean. No model shows a collapse of the MOC by the year 2100.

- Heat waves become more frequent and longer lasting in a future warmer climate. Decreases in frost days are shown to occur almost everywhere in the mid and high latitudes, with an increase in growing season length. There is a tendency for summer drying of the mid-continental areas during summer, indicating a greater risk of droughts in those regions.

- The capacity of the ocean and terrestrial biosphere to absorb anthropogenic CO_2 dioxide is reduced as atmospheric CO_2 increases in the 21st century. This feedback leads to a reduction of the emissions required to achieve a given atmospheric CO_2 stabilization level. The higher the stabilization scenario, the larger the amount of climate change and the larger the impact on the carbon cycle.

Key Uncertainties:

- The likelihood of a large abrupt change of the MOC beyond the end of the 21st century cannot yet be assessed reliably. For low and medium emission scenarios with atmospheric greenhouse gas concentrations stabilized beyond 2100, the MOC recovers from initial weakening within one to several centuries. A permanent reduction of the MOC cannot be excluded if the forcing is strong and long enough.

- The model projections for extremes of precipitation or storms show large ranges in amplitude and geographical locations.

- The response of some major modes of climate variability such as ENSO still differs from model to model, due to differences in the spatial and temporal representation of this natural variability for present-day conditions.

- The robustness of model responses of tropical cyclones, and midlatitude storms, is still limited by the resolution of available climate models.

- Changes to key processes which drive some global and regional climate changes are poorly known (e.g.,. ENSO, NAO, blocking, THC, land-surface feedbacks, tropical cyclone distribution).

- The magnitude of future carbon cycle feedbacks is still poorly determined.

TS.6.4.5 Regional Projections

Robust Findings:

- Over most land areas warming will very likely occur at greater than the global average rate.

- Annual precipitation will very likely increase in most of northern Europe, northern North America, the Arctic, south eastern South America, and the west of the South Island of New Zealand. Winter (DJF) precipitation will very likely increase in Northern Asia, East Asia and the Tibetan Plateau.

- Annual precipitation will very likely decrease in regions of North Africa, northern Sahara and most of the Mediterranean area. Winter rainfall will very likely decrease in much of south western Africa and southern Australia.

- Extremes in daily precipitation will very likely increase in northern Europe, South Asia, East Asia, Southeast Asia, Australia and New Zealand.

Key Uncertainties:

- In some regions there has been only very limited study of key aspects of regional climate change, particularly with regard to extreme events.

- GCMs show no consistency in simulated regional precipitation change in some key regions (e.g., northern South America, northern Australia, the Sahel).

- In many regions there is insufficient information on how climate change will be expressed at fine spatial scale.