Contents

TECHNICAL SUMMARY

Prepared by the lead authors with extensive input from contributors and other experts, from governments and from non-governmental organisations.

This summary, part of the background material on which the Summary for Policymakers is based, has been accepted by the IPCC but not approved in detail. Acceptance by the IPCC signifies the Panel's view that the summary presents a comprehensive, objective and balanced view of the subject matter, but does not confer on this summary the status of a formally-agreed statement by the IPCC.

A Introduction

The IPCC Scientific Assessment Working Group (WGI) was established in 1988 to assess available information on the science of climate change, in particular that arising from human activities. In performing its assessments the Working Group is concerned with:

- developments in the scientific understanding of past and present climate, of climate variability, of climate predictability and of climate change including feedbacks from climate impacts;
- progress in the modelling and projection of global and regional climate and sea level change;
- observations of climate, including past climates, and assessment of trends and anomalies;
- gaps and uncertainties in current knowledge.

The first Scientific Assessment in 1990 concluded that the increase in atmospheric concentrations of greenhouse gases since the pre-industrial period[1] had altered the energy balance of the Earth/atmosphere and that global warming would result. Model simulations of global warming due to the observed increase of greenhouse gas concentrations over the past century tended towards a central estimate of about 1°C while analysis of the instrumental temperature record, on the other hand, revealed warming of around 0.5°C over the same period. The 1990 report concluded: "The size of this warming is broadly consistent with predictions of climate models, but it is also of the same magnitude as natural climate variability. Thus the observed increase could be largely due to this natural variability; alternatively this variability and other human factors could have offset a still larger human-induced greenhouse warming."

A primary concern identified by IPCC (1990) was the expected continued increase in greenhouse gas concentrations as a result of human activity, leading to significant climate change in the coming century. The projected changes in temperature, precipitation and soil moisture were not uniform over the globe. Anthropogenic aerosols were recognised as a possible source of regional cooling but no quantitative estimates of their effects were available.

The IPCC Supplementary Report in 1992 confirmed, or found no reason to alter, the major conclusions of IPCC (1990) . It presented a new range of global mean temperature projections based on a new set of IPCC emission scenarios (IS92 a to f) and reported progress in quantifying the effects of anthropogenic aerosols. Ozone depletion due to chlorofluorocarbons (CFCs) was recognised as a cause of negative radiative forcing, reducing the global importance of CFCs as greenhouse gases.

The 1994 WGI report on Radiative Forcing of Climate Change provided a detailed assessment of the global carbon cycle and of aspects of atmospheric chemistry governing the abundance of non-CO_2 greenhouse gases. Some pathways that would stabilise atmospheric greenhouse gas concentrations were examined, and new or revised calculations of Global Warming Potential for 38 species were presented. The growing literature on processes governing the abundance and radiative properties of aerosols was examined in considerable detail, including new information on the climatic impact of the 1991 eruption of Mt. Pinatubo.

[1] The pre-industrial period is defined as the several centuries preceding 1750.

The Second IPCC Assessment of the Science of Climate Change presents a comprehensive assessment of climate change science as of 1995, including updates of relevant material in all three preceding reports. Key issues examined in the Second Assessment concern the relative magnitude of human and natural factors in driving changes in climate, including the role of aerosols; whether a human influence on present-day climate can be detected; and the estimation of future climate and sea level change at both global and continental scales.

The United Nations Framework Convention on Climate Change (FCCC) uses the term "climate change" to refer exclusively to change brought about by human activities. A more generic usage is common in the scientific community where it is necessary to be able to refer to change arising from any source. In particular scientists refer to past climate change and address the complex issue of separating natural and human causes in currently observed changes. However, the climate projections covered in this document relate only to future climate changes resulting from human influences, since it is not yet possible to predict the fluctuations due to volcanoes and other natural influences. Consequently the use of the term "climate change" here, when referring to future change, is essentially the same as the usage adopted in the FCCC.

Box 1: What drives changes in climate?

The Earth absorbs radiation from the Sun, mainly at the surface. This energy is then redistributed by the atmospheric and oceanic circulation and radiated to space at longer ("terrestrial" or "infrared") wavelengths. On average, for the Earth as a whole, the incoming solar energy is balanced by outgoing terrestrial radiation.

Any factor which alters the radiation received from the Sun or lost to space, or which alters the redistribution of energy within the atmosphere, and between the atmosphere, land and ocean, can affect climate. A *change* in the energy available to the global Earth/atmosphere system is termed here, and in previous IPCC reports, a *radiative forcing*.

Increases in the concentrations of greenhouse gases will reduce the efficiency with which the Earth cools to space. More of the outgoing terrestrial radiation from the surface is absorbed by the atmosphere and emitted at higher altitudes and colder temperatures. This results in a positive radiative forcing which tends to warm the lower atmosphere and surface. This is the *enhanced* greenhouse effect - an enhancement of an effect which has operated in the Earth's atmosphere for billions of years due to the naturally occurring greenhouse gases: water vapour, carbon dioxide, ozone, methane and nitrous oxide. The amount of warming depends on the size of the increase in concentration of each greenhouse gas, the radiative properties of the gases involved, and the concentrations of other greenhouse gases already present in the atmosphere.

Anthropogenic aerosols (small particles) in the troposphere, derived mainly from the emission of sulphur dioxide from fossil fuel burning, and derived from other sources, such as biomass burning, can absorb and reflect solar radiation. In addition, changes in aerosol concentrations can alter cloud amount and cloud reflectivity through their effect on cloud properties. In most cases tropospheric aerosols tend to produce a negative radiative forcing and cool climate. They have a much shorter lifetime (days to weeks) than most greenhouse gases (decades to centuries) so their concentrations respond much more quickly to changes in emissions.

Volcanic activity can inject large amounts of sulphur-containing gases (primarily sulphur dioxide) into the stratosphere which are transformed into aerosols. This can produce a large, but transitory (i.e., a few years), negative radiative forcing, tending to cool the Earth's surface and lower atmosphere over periods of a few years.

The Sun's output of energy varies by small amounts (0.1%) over an 11-year cycle, and variations over longer periods occur. On time-scales of tens to thousands of years, slow variations in the Earth's orbit, which are well understood, have led to changes in the seasonal and latitudinal distribution of solar radiation; these changes have played an important part in controlling the variations of climate in the distant past, such as the glacial cycles.

Any changes in the radiative balance of the Earth, including those due to an increase in greenhouse gases or in aerosols, will tend to alter atmospheric and oceanic temperatures and the associated circulation and weather patterns. These will be accompanied by changes in the hydrological cycle (for example, altered cloud distributions or changes in rainfall and evaporation regimes).

Any human-induced changes in climate will be superimposed on a background of natural climatic variations which occur on a whole range of space- and time-scales. Natural climate variability can occur as a result of changes in the forcing of the climate system, for example due to aerosol derived from volcanic eruptions. Climate variations can also occur in the absence of a change in external forcing, as a result of complex interactions between components of the climate system such as the atmosphere and ocean. The El Niño-Southern Oscillation (ENSO) phenomenon is an example of such natural "internal" variability. To distinguish anthropogenic climate changes from natural variations, it is necessary to identify the anthropogenic "signal" against the background "noise" of natural climate variability.

B Greenhouse gases, aerosols and their radiative forcing

Human activities are changing the atmospheric concentrations and distributions of greenhouse gases and aerosols. These changes can produce a radiative forcing by changing either the reflection or absorption of solar radiation, or the emission and absorption of terrestrial radiation (see Box 1).

Information on radiative forcing was extensively reviewed in IPCC (1994). Summaries of the information in that report and new results are presented here. The most significant advance since IPCC (1994) is improved understanding of the role of aerosols and their representation in climate models.

B.1 Carbon dioxide (CO_2)

CO_2 concentrations have increased from about 280 ppmv in pre-industrial times to 358 ppmv in 1994 (Table 1, and Figure 1a). There is no doubt that this increase is largely due to human activities, in particular fossil fuel combustion, but also land-use conversion and to a lesser extent cement production (Table 2). The increase has led to a radiative forcing of about +1.6 Wm^{-2} (Figure 2).

Prior to this recent increase, CO_2 concentrations over the past 1,000 years, a period when global climate was relatively stable, fluctuated by about ±10 ppmv around 280 ppmv.

Table 1 A sample of greenhouse gases affected by human activities.

	CO_2	CH_4	N_2O	CFC-11	HCFC-22 (a CFC substitute)	CF_4 (a perfluoro-carbon)
Pre-industrial concentration	~280 ppmv	~700 ppbv	~275 ppbv	zero	zero	zero
Concentration in 1994	358 ppmv	1,720 ppbv	312§ ppbv	268§ pptv†	110 pptv	72§ pptv
Rate of concentration change*	1.5 ppmv/yr 0.4%/yr	10 ppbv/yr 0.6%/yr	0.8 ppbv/yr 0.25%/yr	0 pptv/yr 0%/yr	5 pptv/yr 5%/yr	1.2 pptv/yr 2%/yr
Atmospheric lifetime (years)	50–200††	12†††	120	50	12	50,000

§ Estimated from 1992-93 data.
† 1 pptv = 1 part per trillion (million million) by volume.
†† No single lifetime for CO_2 can be defined because of the different rates of uptake by different sink processes.
††† This has been defined as an adjustment time which takes into account the indirect effect of methane on its own lifetime.
*The growth rates of CO_2, CH_4 and N_2O are averaged over the decade beginning 1984 (see Figures 1 and 3); halocarbon growth rates are based on recent years (1990s).

Table 2 Annual average anthropogenic carbon budget for 1980 to 1989. CO_2 sources, sinks and storage in the atmosphere are expressed in GtC/yr.

CO_2 sources	
(1) Emissions from fossil fuel combustion and cement production	5.5 ± 0.5*
(2) Net emissions from changes in tropical land-use	1.6 ± 1.0@
(3) Total anthropogenic emissions = (1)+(2)	7.1 ± 1.1
Partitioning amongst reservoirs	
(4) Storage in the atmosphere	3.3 ± 0.2
(5) Ocean uptake	2.0 ± 0.8
(6) Uptake by Northern Hemisphere forest regrowth	0.5 ± 0.5#
(7) Inferred sink: 3-(4+5+6)	1.3 ± 1.5§

Notes:
* For comparison, emissions in 1994 were 6.1 GtC/yr.
@ Consistent with Chapter 24 of IPCC WGII (1995).
\# This number is consistent with the independent estimate, given in IPCC WGII (1995), of 0.7±0.2 GtC/yr for the mid- and high latitude forest sink.
§ This inferred sink is consistent with independent estimates, given in Chapter9 , of carbon uptake due to nitrogen fertilisation (0.5±1.0 GtC/yr), plus the range of other uptakes (0-2 GtC/yr) due to CO_2 fertilisation and climatic effects.

Figure 1a
(B&W; Postscript file; 2 columns wide)
Figure 1b
(B&W; Postscript file; 1 column wide)
Figure 2
(B&W; Postscript file; 1 column wide)

The annual growth rate of atmospheric CO_2 concentration was low during the early 1990s (0.6 ppmv/yr in 1991/92). However, recent data indicate that the growth rate is currently comparable to that averaged over the 1980s, around 1.5 ppmv/yr (Figure 1b). Isotopic data suggest that the low growth rate resulted from fluctuations in the exchanges of CO_2 between the atmosphere and both the ocean and the terrestrial biosphere, possibly resulting from climatic and biospheric variations following the eruption of Mt. Pinatubo in June 1991. While understanding these short-term fluctuations is important, fluctuations of a few years' duration are not relevant to projections of future concentrations or emissions aimed at estimating longer time-scale changes to the climate system.

The estimate of the 1980s' carbon budget (Table 2) remains essentially unchanged from IPCC (1994). While recent data on anthropogenic emissions are available, there are insufficient analyses of the other fluxes to allow an update of this decadal budget to include the early years of the 1990s. The net release of carbon from tropical land-use change (mainly forest clearing minus regrowth) is roughly balanced by carbon accumulation in other land ecosystems due to forest regrowth outside the tropics, and by transfer to other reservoirs stimulated by CO_2 and nitrogen fertilisation and by decadal time-scale climatic effects. Model results suggest that during the 1980s, CO_2 fertilisation resulted in a transfer of carbon from the atmosphere to the biosphere of 0.5 to 2.0 GtC/yr and nitrogen fertilisation resulted in a transfer of carbon from the atmosphere to the biosphere of between 0.2 and 1.0 GtC/yr.

CO_2 is removed from the atmosphere by a number of processes that operate on different time-scales, and is subsequently transferred to various reservoirs, some of which eventually return CO_2 to the atmosphere. Some simple analyses of CO_2 changes have used the concept of a single characteristic time-scale for this gas. Such analyses are of limited value because a single time-scale cannot capture the behaviour of CO_2 under different emission scenarios. This is in contrast to methane, for example, whose atmospheric lifetime is dominantly controlled by a single process : oxidation by OH in the atmosphere. For CO_2 the fastest process is uptake into vegetation and the surface layer of the oceans which occurs over a few years. Various other sinks operate on the century time-scale (e.g., transfer to soils and to the deep ocean) and so have a less immediate, but no less important, effect on the atmospheric concentration. Within 30 years about 40-60% of the CO_2 currently released to the atmosphere is removed. However, if emissions were reduced, the CO_2 in the vegetation and ocean surface water would soon equilibrate with that in the atmosphere, and the rate of removal would then be determined by the slower response of woody vegetation, soils, and transfer into the deeper layers of the ocean. Consequently, most of the excess atmospheric CO_2 would be removed over about a century although a portion would remain airborne for thousands of years because transfer to the ultimate sink - ocean sediments - is very slow.

There is large uncertainty associated with the future role of the terrestrial biosphere in the global carbon budget for several reasons. First, future rates of deforestation and regrowth in the tropics and mid-latitudes are difficult to predict. Second, mechanisms such as CO_2 fertilisation remain poorly quantified at the ecosystem level. Over decades to centuries, anthropogenic changes in atmospheric CO_2 content and climate may also alter the global distribution of ecosystem types. Carbon could be released rapidly from areas where forests die, although regrowth could eventually sequester much of this carbon. Estimates of this loss range from near zero to, at low probabilities, as much as 200 GtC over the next one-to-two centuries, depending on the rate of climate change.

The marine biota both respond to and can influence climate change. Marine biota play a critical role in depressing the atmospheric CO_2 concentration significantly below its equilibrium state in the absence of biota. Changes in nutrient supply to the surface ocean resulting from changes in ocean circulation, coastal runoff and atmospheric deposition, and changes in the amount of sea ice and cloudiness, have the potential to affect marine biogeochemical processes. Such changes would be expected to have an impact (at present unquantifiable) on the cycling of CO_2 and the production of other climatically important trace gases. It has been suggested that a lack of iron limits phytoplankton growth in certain ocean areas. However, it is not likely that iron fertilisation of CO_2 uptake by phytoplankton can be used to draw down atmospheric CO_2: even massive continual seeding of 10-15% of the world oceans (the Southern Ocean) until 2100, if it worked with 100% efficiency and no opposing side-effects (e.g., increased N_2O production), would reduce the atmospheric CO_2 build-up projected by the IPCC (1990) "Business-as-usual" emission scenario by less than 10%.

B.2 Methane (CH_4)

Methane is another naturally occurring greenhouse gas whose concentration in the atmosphere is growing as a result of human activities such as agriculture and waste disposal, and fossil fuel production and use (Table 3).

Table 3 Estimated sources and sinks of methane for 1980 to 1990. All figures are in Tg† (CH_4)/yr. The current global atmospheric burden of CH_4 is about 5000 Tg(CH_4).

(a) Observed atmospheric increase, estimated sinks and sources derived to balance the budget .

		Individual estimates	Total
Atmospheric increase*			**37** (35-40)
Sinks of atmospheric CH_4:	tropospheric OH	**490** (405-575)	
	stratosphere	**40** (32–48)	
	soils	**30** (15–45)	
Total atmospheric sinks			**560** (460-660)
Implied sources (sinks + atmospheric increase)			**597** (495–700)

(b) Inventory of identified sources.

		Individual estimates	Total
Natural sources			**160** (110–210)
Anthropogenic sources:	Fossil fuel related	**100** (70–120)	
	Total biospheric	**275** (200–350)	
Total anthropogenic sources			**375** (300–450)
Total identified sources			**535** (410–660)

† 1Tg = 1 million million grams, which is equivalent to 1 million tonnes.
* Applies to 1980-1990 average. Table 1 and the stabilisation discussion in Section B.9.2 use the average for 1984-1994.

Global average methane concentrations increased by 6% over the decade starting in 1984 (Figure 3). Its concentration in 1994 was about 1720 ppbv, 145% greater than the pre-industrial concentration of 700 ppbv (Table 1, Figure 3). Over the last 20 years, there has been a decline in the methane growth rate: in the late 1970s the concentration was increasing by about 20 ppbv/yr, during the 1980s the growth rate dropped to 9-13 ppbv/yr. Around the middle of 1992, methane concentrations briefly stopped growing, but since 1993 the global growth rate has returned to about 8 ppbv/yr.

Figure 3
(B&W; Postscript file; 1 column wide)

Individual methane sources are not well quantified. Carbon isotope measurements indicate that about 20% of the total annual methane emissions are related to the production and use of fossil fuel. In total, anthropogenic activities are responsible for about 60-80% of current methane emissions (Table 3). Methane emissions from natural wetlands appear to contribute about 20% to the global methane emissions to the atmosphere. Such emissions will probably increase with global warming as a result of greater microbial activity. In 1992 the direct radiative forcing due to the increase in methane concentration since pre-industrial times was about +0.47 Wm^{-2} (Figure 2).

Changes in the concentration of methane have clearly-identified chemical feedbacks. The main removal process for methane is reaction with the hydroxyl radical (OH). Addition of methane to the atmosphere reduces the concentration of tropospheric OH which can in turn feed back and reduce the rate of methane removal.

The adjustment time for a pulse of methane added to the atmosphere has been revised to 12 (±3) years (compared with 14.5 (±2.5) years in IPCC (1994)). Two factors are responsible for the change: (a) a new estimate for the chemical removal rate (11% faster); and (b) inclusion of the uptake of methane by soils. The revised global sink strength is 560 (±100) $Tg(CH_4)$/year, higher than the 1994 estimate, but still consistent with the previous range of global source strength.

B.3 Nitrous oxide (N_2O)

There are many small sources of nitrous oxide, both natural and anthropogenic, which are difficult to quantify. The main anthropogenic sources are from agriculture and a number of industrial processes (e.g., adipic acid and nitric acid production). A best estimate of the current (1980s) anthropogenic emission of nitrous oxide is 3 to 8 Tg(N)/yr. Natural sources are poorly quantified, but are probably twice as large as anthropogenic sources. Nitrous oxide is removed mainly by photolysis (breakdown by sunlight) in the stratosphere and consequently has a long lifetime (about 120 years).

Although sources cannot be well quantified, atmospheric measurements and evidence from ice cores show that the atmospheric abundance of nitrous oxide has increased since the pre-industrial era, most likely owing to human activities. In 1994 atmospheric levels of nitrous oxide were about 312 ppbv; pre-industrial levels were about 275 ppbv (Table 1). The 1993 growth rate (approximately 0.5 ppbv/yr) was lower than that observed in the late 1980s and early 1990s (approximately 0.8 ppbv/yr), but these short-term changes in growth rate are within the range of variability seen on decadal time-scales. The radiative forcing due to the change in nitrous oxide since pre-industrial times is about +0.14 Wm^{-2} (Figure 2).

B.4 Halocarbons and other halogenated compounds

Halocarbons are carbon compounds containing fluorine, chlorine, bromine or iodine. Many of these are effective greenhouse gases. For most of these compounds, human activities are the sole source.

Halocarbons that contain chlorine (CFCs and HCFCs) and bromine (halons) cause ozone depletion, and their emissions are controlled under the Montreal Protocol and its Adjustments and Amendments. As a result, growth rates in the concentrations of many of these compounds have already fallen (Figure 4) and the radiative impact of these compounds will slowly decline over the next century. The contribution to *direct* radiative forcing due to concentration increases of these CFCs and HCFCs since pre-industrial times is about +0.25 Wm^{-2}. Halocarbons can also exert an *indirect* negative radiative forcing through their depletion of stratospheric ozone (see Section B.5.2).

Figure 4
(B&W; Postscript file; 1 column wide)

Perfluorocarbons (PFCs, e.g., CF_4, C_2F_6) and sulphur hexafluoride (SF_6) are removed very slowly from the atmosphere with estimated lifetimes greater than 1000 years. As a result, effectively all emissions accumulate in the atmosphere and will continue to influence climate for thousands of years. Although the radiative forcing due to concentration increases of these compounds since pre-industrial times is small (about +0.01 Wm^{-2}), it may become significant in the future if concentrations continue to increase.

Hydrofluorocarbons (HFCs) are being used to replace ozone-depleting substances in some applications; their concentrations and radiative impacts are currently small. If emissions increase as envisaged in Scenario IS92a , they would contribute about 3% of the total radiative forcing from all greenhouse gases by the year 2100.

B.5 Ozone (O_3)

Ozone is an important greenhouse gas present in both the stratosphere and troposphere. Changes in ozone cause radiative forcing by influencing both solar and terrestrial radiation. The net radiative forcing is strongly dependent on the vertical distribution of ozone change and is particularly sensitive to changes around the tropopause level, where trends are difficult to estimate due to a lack of reliable observations and the very large natural variability. The patterns of both tropospheric and stratospheric ozone changes are spatially variable. Estimation of the radiative forcing due to changes in ozone is thus more complex than for the well-mixed greenhouse gases.

B.5.1 Tropospheric Ozone

In the troposphere, ozone is produced during the oxidation of methane and from various short-lived precursor gases (mainly carbon monoxide (CO), nitrogen oxides (NO_x) and non-methane hydrocarbons (NMHC)). Ozone is also transported into the troposphere from the stratosphere. Changes in tropospheric ozone concentration are spatially variable, both regionally and vertically, making assessment of global long-term trends difficult. In the Northern Hemisphere, there is some evidence that tropospheric ozone concentrations have increased since 1900, with strong evidence that this has occurred in many locations since the 1960s. However, the observations of the most recent decade show that the upward trend has slowed significantly or stopped. Model simulations and the limited observations together suggest that ozone concentrations throughout the troposphere may have doubled in the Northern Hemisphere since pre-industrial times, an increase of about 25 ppbv. In the Southern Hemisphere, there are insufficient data to determine if tropospheric ozone has changed, except at the South Pole where a decrease has been observed since the mid-1980s.

Changes in tropospheric ozone have potentially important consequences for radiative forcing. The calculated global average radiative forcing due to the increased concentration since pre-industrial times is +0.4 (±0.2) Wm^{-2}.

B.5.2 Stratospheric Ozone

Decreases in stratospheric ozone have occurred since the 1970s, principally in the lower stratosphere. The most obvious feature is the annual appearance of the Antarctic "ozone hole" in September and October. The October average total ozone values over Antarctica are 50–70% lower than those observed in the 1960s. Statistically significant losses in total ozone have also been observed in the mid-latitudes of both hemispheres. Little or no downward trend in ozone has been observed in the tropics (20°N–20°S). The weight of recent scientific evidence strengthens the previous conclusion that ozone loss is due largely to anthropogenic chlorine and bromine compounds. Since the stratospheric abundances of chlorine and bromine are expected to continue to grow for a few more years before they decline (see Section B.4), stratospheric ozone losses are expected to peak near the end of the century, with a gradual recovery throughout the first half of the 21st century.

The loss of ozone in the lower stratosphere over the past 15 to 20 years has led to a globally averaged radiative forcing of about -0.1 Wm^{-2}. This negative radiative forcing represents an indirect effect of anthropogenic chlorine and bromine compounds.

B.6 Tropospheric and stratospheric aerosols

Aerosol is a term used for particles and very small droplets of natural and human origin that occur in the atmosphere; they include dust and other particles which can be made up of many different chemicals. Aerosols are produced by a variety of processes, both natural (including dust storms and volcanic activity) and anthropogenic (including fossil fuel and biomass burning). Aerosols contribute to visible haze and can cause a diminution of the intensity of sunlight at the ground.

Aerosols in the atmosphere influence the radiation balance of the Earth in two ways: (i) by scattering and absorbing radiation - the *direct* effect, and (ii) by modifying the optical properties, amount and lifetime of clouds - the *indirect* effect. Although some aerosols, such as soot, tend to warm the surface, the net climatic effect of anthropogenic aerosols is believed to be a negative radiative forcing, tending to cool the surface (see Section B.7 and Figure 2).

Most aerosols with anthropogenic sources are found in the lower troposphere (below 2 km). Aerosols undergo chemical and physical transformations in the atmosphere, especially within clouds, and are removed largely by precipitation. Consequently aerosols in the lower troposphere typically have residence times of a few days. Because of their short lifetime, aerosols in the lower troposphere are distributed inhomogeneously with maxima close to the natural (especially desert) and anthropogenic (especially industrial and biomass combustion) source regions. Aerosol particles resulting from volcanic activity can reach the stratosphere where they are transported around the globe over many months or years.

The radiative forcing due to aerosols depends on the size, shape and chemical composition of the particles and the spatial distribution of the aerosol. While these factors are comparatively well-known for stratospheric aerosols, there remain many uncertainties concerning tropospheric aerosols.

Since IPCC (1994), there have been several advances in understanding the impact of tropospheric aerosols on climate. These include: (i) new calculations of the spatial distribution of sulphate aerosol largely resulting from fossil fuel combustion and (ii) the first calculation of the spatial distribution of soot aerosol. The impact of these developments on the calculation of aerosol radiative forcing is discussed in Section B.7.

B.7 Summary of radiative forcing

Globally averaged radiative forcing is a useful concept for giving a first-order estimate of the potential climatic importance of various forcing mechanisms. However, as was emphasised in IPCC (1994), there are limits to its utility. In particular, the spatial patterns of forcing differ between the globally well-mixed greenhouse gases, the regionally-varying tropospheric ozone, and the even more regionally concentrated tropospheric aerosols, and so a comparison of the global mean radiative forcings does not give a complete picture of their possible climatic impact.

Estimates of the radiative forcings due to changes in greenhouse gas concentrations since pre-industrial times remain unchanged from IPCC (1994) (see Figure 2). These are +2.45 Wm^{-2} (range: +2.1 to +2.8 Wm^{-2}) for the direct effect of the main well-mixed greenhouse gases (CO_2, CH_4, N_2O and the halocarbons), +0.4 Wm^{-2} (range: 0.2 to 0.6 Wm^{-2}) for tropospheric ozone and -0.1 Wm^{-2} (range: -0.05 to -0.2 Wm^{-2}) for stratospheric ozone.

The total direct forcing due to anthropogenic aerosol (sulphates, fossil fuel soot and organic aerosols from biomass burning) is estimated to be -0.5 Wm^{-2} (range: -0.25 to -1.0 Wm^{-2}). This estimate is smaller than that given in IPCC (1994) owing to a reassessment of the model results used to derive the geographic distribution of aerosol particles and the inclusion of anthropogenic soot aerosol for the first time. The direct forcing due to sulphate aerosols resulting from fossil fuel emissions and smelting is estimated to be -0.4 Wm^{-2} (range: -0.2 to -0.8 Wm^{-2}). The first estimates of the impact of soot in aerosols from fossil fuel sources have been made: significant uncertainty remains but an estimate of +0.1 Wm^{-2} (range: 0.03 to 0.3 Wm^{-2}) is made. The direct radiative forcing since 1850 of particles associated with biomass burning is estimated to be -0.2 Wm^{-2} (range:

-0.07 to -0.6 Wm^{-2}), unchanged from IPCC (1994). It has recently been suggested that a significant fraction of the tropospheric dust aerosol is influenced by human activities but the radiative forcing of this component has not yet been quantified.

The range of estimates for the radiative forcing due to changes in cloud properties caused by aerosols arising from human activity (the indirect effect) is unchanged from IPCC (1994) at between 0 and -1.5 Wm^{-2}. Several new studies confirm that the indirect effect of aerosol may have caused a substantial negative radiative forcing since pre-industrial times, but it remains very difficult to quantify, more so than the direct effect. While no best estimate of the indirect forcing can currently be made, the central value of -0.8 Wm^{-2} has been used in some of the scenario calculations described in Sections B.9.2 and F.2.

There are no significant alterations since IPCC (1994) in the assessment of radiative forcing caused by changes in solar radiative output or stratospheric aerosol loading resulting from volcanic eruptions. The estimate of radiative forcing due to changes in solar radiative output since 1850 is +0.3 Wm^{-2} (range: +0.1 to +0.5). Radiative forcing due to volcanic aerosols resulting from an individual eruption can be large (the maximum global mean effect from the eruption of Mt. Pinatubo was -3 to -4 Wm^{-2}), but lasts for only a few years. However, the transient variations in both these forcings may be important in explaining some of the observed climate variations on decadal time-scales.

B.8 Global Warming Potential (GWP)

The Global Warming Potential is an attempt to provide a simple measure of the relative radiative effects of the emissions of various greenhouse gases. The index is defined as the cumulative radiative forcing between the present and some chosen time horizon caused by a unit mass of gas emitted now, expressed relative to that for some reference gas (here CO_2 is used). The future global warming commitment of a greenhouse gas over a chosen time horizon can be estimated by multiplying the appropriate GWP by the amount of gas emitted. For example, GWPs could be used to compare the effects of reductions in CO_2 emissions relative to reductions in methane emissions, for a specified time horizon.

Derivation of GWPs requires knowledge of the fate of the emitted gas and the radiative forcing due to the amount remaining in the atmosphere. Although the GWPs are quoted as single values, the typical uncertainty is ±35%, not including the uncertainty in the carbon dioxide reference. Because GWPs are based on the radiative forcing concept, they are difficult to apply to radiatively important constituents that are unevenly distributed in the atmosphere. No attempt is made to define a GWP for aerosols. Additionally the choice of time horizon will depend on policy considerations.

GWPs need to take account of any indirect effects of the emitted greenhouse gas, if they are to reflect correctly future warming potential. The net GWPs for the ozone-depleting gases, which include the direct "warming" and indirect "cooling" effects, have now been estimated. In IPCC (1994), only the direct GWPs were presented for these gases. The indirect effect reduces their GWPs, but each ozone-depleting gas must be considered individually. The net GWPs of the chlorofluorocarbons (CFCs) tend to be positive, while those of the halons tend to be negative. The calculation of indirect effects for a number of other gases (e.g., NO_x, CO) is not currently possible because of inadequate characterisation of many of the atmospheric processes involved.

Updates or new GWPs are given for a number of key species (Table 4), based on improved or new estimates of atmospheric lifetimes, molecular radiative forcing factors, and improved representation of the carbon cycle. Revised lifetimes for gases destroyed by chemical reactions in the lower atmosphere (particularly methane, HCFCs and HFCs) have resulted in GWPs that are slightly lower (typically by 10-15%) than those cited in IPCC (1994). The IPCC definition of GWP is based on calculating the relative radiative impact of a release of a trace gas over a time horizon in a constant background atmosphere. In a future atmosphere with larger CO_2 concentrations, such as occur in all of the IPCC emission scenarios (see Figure 5b), we would calculate slightly larger GWP values than those given in Table 4.

Table 4 Global Warming Potential referenced to the updated decay response for the Bern carbon cycle model and future CO_2 atmospheric concentrations held constant at current levels.

Species	Chemical Formula	Lifetime (years)	Global Warming Potential (Time Horizon) 20 years	100 years	500 years
CO_2	CO_2	variable§	1	1	1
Methane*	CH_4	12±3	56	21	6.5
Nitrous oxide	N_2O	120	280	310	170
HFC-23	CHF_3	264	9,100	11,700	9,800
HFC-32	CH_2F_2	5.6	2,100	650	200
HFC-41	CH_3F	3.7	490	150	45
HFC-43-10mee	$C_5H_2F_{10}$	17.1	3,000	1,300	400
HFC-125	C_2HF_5	32.6	4,600	2,800	920
HFC-134	$C_2H_2F_4$	10.6	2,900	1,000	310
HFC-134a	CH_2FCF_3	14.6	3,400	1,300	420
HFC-152a	$C_2H_4F_2$	1.5	460	140	42
HFC-143	$C_2H_3F_3$	3.8	1,000	300	94
HFC-143a	$C_2H_3F_3$	48.3	5,000	3,800	1,400
HFC-227ea	C_3HF_7	36.5	4,300	2,900	950
HFC-236fa	$C_3H_2F_6$	209	5,100	6,300	4,700
HFC-245ca	$C_3H_3F_5$	6.6	1,800	560	170
Sulphur hexafluoride	SF_6	3200	16,300	23,900	34,900
Perfluoromethane	CF_4	50000	4,400	6,500	10,000
Perfluoroethane	C_2F_6	10000	6,200	9,200	14,000
Perfluoropropane	C_3F_8	2600	4,800	7,000	10,100
Perfluorobutane	C_4F_{10}	2600	4,800	7,000	10,100
Perfluorocyclo-butane	$c\text{-}C_4F_8$	3200	6,000	8,700	12,700
Perfluoropentane	C_5F_{12}	4100	5,100	7,500	11,000
Perfluorohexane	C_6F_{14}	3200	5,000	7,400	10,700
Ozone-depleting substances†	e.g., CFCs and HCFCS.				

§ Derived from the Bern carbon cycle model.
* The GWP for methane includes indirect effects of tropospheric ozone production and stratospheric water vapour production, as in IPCC (1994). The updated adjustment time for methane is discussed in Section B.2.
† The Global Warming Potentials for ozone-depleting substances (including all CFCs, HCFCs and halons, whose direct GWPs have been given in previous reports) are a sum of a direct (positive) component and an indirect (negative) component which depends strongly upon the effectiveness of each substance for ozone destruction. Generally, the halons are likely to have negative net GWPs, while those of the CFCs are likely to be positive over both 20- and 100-year time horizons (see Chapter 2, Table 2.8).

B.9 Emissions and concentrations of greenhouse gases and aerosols in the future

B.9.1 The IS92 emission scenarios

The projection of future anthropogenic climate change depends, among other things, on assumptions made about future emissions of greenhouse gases and aerosol precursors and the proportion of emissions remaining in the atmosphere. Here we consider the IS92 emission scenarios (IS92a to f) which were first discussed in IPCC (1992).

The IS92 emission scenarios extend to the year 2100 and include emissions of CO_2, CH_4, N_2O, the halocarbons (CFCs and their substitute HCFCs and HFCs), precursors of tropospheric ozone and sulphate aerosols and aerosols from biomass burning. A wide range of assumptions regarding future economic, demographic and policy factors are encompassed (IPCC, 1992). In this report, the emissions of chlorine- and bromine-containing halocarbons listed in IS92 are assumed to be phased out under the Montreal Protocol and its Adjustments and Amendments and so a single revised future emission scenario for these gases is incorporated in all of the IS92 scenarios.

Emissions of individual HFCs are based on the original IS92 scenarios, although they do not reflect current markets. CO_2 emissions for the six scenarios are shown in Figure 5a.

Figure 5a
(B&W; Postscript file; 1 column wide)
Figure 5b
(B&W; Postscript file; 1 column wide)

The calculation of future concentrations of greenhouse gases, given certain emissions, entails modelling the processes that transform and remove the different gases from the atmosphere. For example, future concentrations of CO_2 are calculated using models of the carbon cycle which model the exchanges of CO_2 between the atmosphere and the oceans and terrestrial biosphere (see Section B.1); atmospheric chemistry models are used to simulate the removal of chemically active gases such as methane.

All the IS92 emission scenarios, even IS92c, imply increases in greenhouse gas concentrations from 1990 to 2100 (e.g., CO_2 increases range from 35 to 170% (Figure 5b); CH_4 from 22 to 175%; and N_2O from 26 to 40%).

For greenhouse gases, radiative forcing is dependent on the concentration of the gas and the strength with which it absorbs and re-emits long-wave radiation. For sulphate aerosol, the direct and indirect radiative forcings were calculated on the basis of sulphur emissions contained in the IS92 scenarios. The radiative forcing due to aerosol from biomass burning was assumed to remain constant at -0.2 Wm^{-2} after 1990. The contribution from aerosols is probably the most uncertain part of future radiative forcing.

Figure 6a shows a single "best estimate" of historical radiative forcing from 1765 to 1990 (including the effects of aerosols), followed by radiative forcing for Scenarios IS92 a to f. Figures 6b and c show the contribution to future radiative forcing from various components of the IS92a Scenario; the largest contribution comes from CO_2, with a radiative forcing of almost +6 Wm^{-2} by 2100. The negative forcing due to tropospheric aerosols, in a globally averaged sense, offsets some of the greenhouse gas positive forcing. However, because tropospheric aerosols are highly variable regionally, their globally averaged radiative forcing will not adequately describe their possible climatic impact.

Figure 6a
(B&W; Postscript file; 1 column wide)
Figure 6b
(B&W; Postscript file; 1 column wide)
Figure 6c
(B&W; Postscript file; 1 column wide)

Future projections of temperature and sea level based on the IS92 emissions scenarios are discussed in Section F.

B.9.2 Stabilisation of greenhouse gas and aerosol concentrations

An important question to consider is: how might greenhouse gas concentrations be stabilised in the future?

If global CO_2 emissions were maintained at near current (1994) levels, they would lead to a nearly constant rate of increase in atmospheric concentrations for at least two centuries, reaching about 500 ppmv (approaching twice the pre-industrial concentration of 280 ppmv) by the end of the 21st century.

In IPCC (1994), carbon cycle models were used to calculate the emissions of CO_2 which would lead to stabilisation at a number of different concentration levels from 350 to 750 ppmv. The assumed concentration profiles leading to stabilisation are shown in Figure 7a (excluding 350 ppmv). Many different stabilisation levels, time-scales for achieving these levels, and routes to stabilisation could have been chosen. The choices made are not intended to have policy implications; the exercise is illustrative of the relationship between CO_2 emissions and concentrations. Those in Figure 7a assume a smooth transition from the current average rate of CO_2 concentration increase to stabilisation. To a first approximation, the stabilised concentration level depends more upon the accumulated amount of carbon emitted up to the time of stabilisation, than upon the exact concentration path followed *en route* to stabilisation.

Figure 7a
(B&W; Postscript file; 1 column wide)
Figure 7b
(B&W; Postscript file; 1 column wide)

New results have been produced to take account of the revised carbon budget for the 1980s (Table 2), but the main conclusion, that stabilisation of concentration requires emissions eventually to drop well below current levels, remains unchanged from IPCC (1994) (Figure 7b). Because the new budget implies a reduced terrestrial sink, the allowable emissions to achieve stabilisation are up to 10% lower than those in IPCC (1994). In addition, these calculations have been extended to include alternative pathways towards stabilisation (Figure 7a) and a higher stabilisation level (1000 ppmv). The alternative pathways assume higher emissions in the early years, but require steeper reductions in emissions in later years (Figure 7b). The 1000 ppmv stabilisation case allows higher maximum emissions, but still requires a decline to current levels by about 240 years from now and further reductions thereafter (Figure 7b).

The accumulated anthropogenic CO_2 emissions from 1991 to 2100 inclusive are shown in Table 5 for the profiles leading to stabilisation at 450, 550, 650, 750 and 1000 ppmv via the profiles shown in Figure 7a and, for comparison, the IS92 emission scenarios. These values are calculated using the Bern carbon cycle model. Based on the results in IPCC (1994) it is estimated that values calculated with different carbon cycle models could be up to approximately 15% higher or lower than those presented here.

Table 5 Total anthropogenic CO_2 emissions accumulated from 1991 to 2100 inclusive (GtC). All values were calculated using the carbon budget for 1980s shown in Table 2 and the Bern carbon cycle model.

Case		Accumulated CO_2 emissions 1991 to 2100 (GtC)	
IS92 scenarios	c	770	
	d	980	
	b	1430	
	a	1500	
	f	1830	
	e	2190	
		Concentration profiles A*	**Concentration profiles B†**
Stabilisation at	450 ppmv	630	650
	550 ppmv	870	990
	650 ppmv	1030	1190
	750 ppmv	1200	1300
	1000 ppmv	-	1410

* As in IPCC (1994) - see Figure 7a.
† Profiles that allow emissions to follow IS92a until at least the year 2000 - see Figure 7a.

If methane emissions were to remain constant at 1984-1994 levels (i.e., those sustaining an atmospheric trend of +10 ppbv/yr), the methane concentration would rise to about 1820 ppbv over the next 40 years. If emissions were cut by about 30 $Tg(CH_4)$/yr (about 8% of current anthropogenic emissions), CH_4 concentrations would remain at today's levels. These estimates are lower than those in IPCC (1994).

If emissions of N_2O were held constant at today's level, the concentration would climb from 312 ppbv to about 400 ppbv over several hundred years. In order for the concentration to be stabilised near current levels, anthropogenic sources would need to be reduced by more than 50%. Stabilisation of PFCs and SF_6 concentrations can only be achieved effectively by stopping emissions.

Because of their short lifetime, future tropospheric aerosol concentrations would respond almost immediately to changes in emissions. For example, control of sulphur emissions would immediately reduce the amount of sulphate aerosol in the atmosphere.

C Observed trends and patterns in climate and sea level

Section B demonstrated that human activities have changed the concentrations and distributions of greenhouse gases and aerosols over the 20th century; this section discusses the changes in temperature, precipitation (and related hydrological variables), climate variability and sea level that have been observed over the same period. Whether the observed changes are in part induced by human activities is considered in Section E.

C.1 Has the climate warmed?

Global average surface air temperature, excluding Antarctica, is about 15°C. Year-to-year temperature *changes* can be computed with much more confidence than the absolute global average temperature.

The mean global surface temperature has increased by about 0.3° to 0.6°C since the late 19th century, and by about 0.2° to 0.3°C over the last 40 years , the period with most credible data (see Figure 8 which shows data up to the end of 1994). The warming occurred largely during two periods, between 1910 and 1940 and since the mid-1970s. The estimate of warming has not significantly changed since the IPCC (1990) and IPCC (1992). Warming is evident in both sea surface and land-based surface air temperatures. Urbanisation in general and desertification could have contributed only a small part (a few hundredths of a degree) of the overall global warming, although urbanisation influences may have been important in some regions. Indirect indicators, such as borehole temperatures and glacier shrinkage, provide independent support for the observed warming. Recent years have been among the warmest since 1860, i.e., in the period of instrumental record.

Figure 8
(B&W; Postscript file; 1 column wide)

The warming has not been globally uniform. The recent warmth has been greatest over the continents between 40°N and 70°N. A few areas, such as the North Atlantic Ocean north of 30°N, and some surrounding land areas, have cooled in recent decades (Figure 9).

Figure 9
(Colour; Postscript file; 2 column wide)

As predicted in IPCC (1992) and discussed in IPCC (1994), relatively cooler global surface and tropospheric temperatures (by about 0.5°C) and a relatively warmer lower stratosphere (by about 1.5°C) were observed in 1992 and early 1993, following the 1991 eruption of Mt. Pinatubo. Warmer temperatures at the surface and in the lower troposphere, and a cooler lower stratosphere, reappeared in 1994 following the removal by natural processes of Mt. Pinatubo aerosols from the stratosphere.

The general tendency toward reduced daily temperature range over land, at least since the middle of the 20th century, noted in IPCC (1992), has been confirmed with more data (which have now been analysed for more than 40% of the global land area). The range has decreased in many areas because nights have warmed more than days. Minimum temperatures have typically increased twice as much as maximum temperatures over the last 40 years. A likely explanation, in addition to the effects of enhanced greenhouse gases, is an increase in cloud cover which has been observed in many of the areas with reduced diurnal temperature range. An increase in cloud reduces diurnal temperature range both by obstructing daytime sunshine, and by preventing the escape of terrestrial radiation at night. Anthropogenic aerosols may also have an influence on daily temperature range.

Temperature trends in the free atmosphere are more difficult to determine than at the surface as there are fewer data and the records are much shorter. Radiosonde data which are available since the 1950s show warming trends of around 0.1°C per decade, as at the surface, but since 1979 when satellite data of global tropospheric temperatures became available, there appears to have been a slight cooling (about -0.06°C per decade), whereas surface measurements still show a warming. These apparently contradictory trends can be reconciled if the diverse response of the troposphere and surface to short-term events such as volcanic eruptions and El Niño are taken into account. After adjustment for these transient effects, both tropospheric and surface data show slight warming (about 0.1°C per decade for the troposphere and nearly 0.2°C per decade at the surface) since 1979.

Cooling of the lower stratosphere has been observed since 1979 both by satellites and weather balloons, as noted in IPCC (1992) and IPCC (1994). Current global mean stratospheric temperatures are the coldest observed in the relatively short period of the record. Reduced stratospheric temperature has been projected to accompany both ozone losses in the lower stratosphere and atmospheric increases of carbon dioxide.

C.2 Is the 20th century warming unusual?

In order to establish whether the 20th century warming is part of the natural variability of the climate system or a response to anthropogenic forcing, information is needed on climate variability on relevant time-scales. As an average over the Northern Hemisphere for summer, recent decades appear to be the warmest since at least 1400 from the limited available evidence (Figure 10). The warming over the past century began during one of the colder periods of the last 600 years. Prior to 1400 data are insufficient to provide hemispheric temperature estimates. Ice core data from several sites suggest that the 20th century is at least as warm as any century in the past 600 years, although the recent warming is not exceptional everywhere.

Figure 10
(B&W; Postscript file; 1 column wide)

Large and rapid climatic changes occurred during the last glacial period (around 20,000 to 100,000 years ago) and during the transition period towards the present interglacial (the last 10,000 years, known as the Holocene). Changes in annual mean temperature of about 5°C occurred over a few decades, at least in Greenland and the North Atlantic, and were probably linked to changes in oceanic circulation. These rapid changes suggest that climate may be quite sensitive to internal or external climate forcings and feedbacks. The possible relevance of these rapid climate changes to future climate is discussed in Section F.5.

Temperatures have been less variable during the last 10,000 years. Based on the incomplete evidence available, it is unlikely that global mean temperatures have varied by more than 1°C in a century during this period.

C.3 Has the climate become wetter?

As noted in IPCC (1992), precipitation has increased over land in high latitudes of the Northern Hemisphere, especially during the cold season. A step-like decrease of precipitation occurred after the 1960s over the subtropics and tropics from Africa to Indonesia. These changes are consistent with changes in streamflow, lake levels and soil moisture (where data analyses are available). Precipitation, averaged over global land areas, increased from the start of the century up to about 1960. Since about 1980 precipitation over land has decreased (Figure 11).

Figure 11
(B&W; Postscript file; 1 column wide)

There is evidence to suggest increased precipitation over the central equatorial Pacific Ocean in recent decades, with decreases to the north and south. Lack of data prevents us from reaching firm conclusions about other precipitation changes over the ocean.

Estimates suggest that evaporation may have increased over the tropical oceans (although not everywhere) but decreased over large portions of Asia and North America. There has also been an observed increase in atmospheric water vapour in the tropics, at least since 1973.

Cloudiness appears to have increased since the 1950s over the oceans. In many land areas where the daily temperature range has decreased (see Section C.1), cloudiness increased from the 1950s to at least the 1970s.

Snow cover extent over the Northern Hemisphere land surface has been consistently below the 21-year average (1974 to 1994) since 1988. Snow-radiation feedback has amplified springtime warming over mid- to high latitude Northern Hemisphere land areas.

A summary of observed climate trends is shown in Figure 12.

Figure 12
(B&W; Postscript file; 2 columns wide)

C.4 Has sea level risen?

Over the last 100 years global sea level has risen by about 10 to 25 cm, based on analyses of tide-gauge records. A major source of uncertainty in estimating the rate of rise is the influence of vertical land movements, which are included in sea level measurements made by tide gauges. Since IPCC (1990), improved methods for filtering out the effects of long-term vertical land movements, as well as greater reliance on the longest tide-gauge records for estimating trends, have provided greater confidence that the volume of ocean water has, in fact, been increasing and causing sea level to rise within the indicated range.

It is likely that much of the rise in sea level has been related to the concurrent rise in the global temperature over the last 100 years. On this time-scale, the warming and consequent expansion of the oceans may account for about 2 to 7 cm of the observed rise in sea level, while the observed retreat of glaciers and ice-caps may account for about 2 to 5 cm. Other factors are more difficult to quantify. Changes in surface and ground water storage may have caused a small change in sea level over the last 100 years. The rate of observed sea level rise suggests that there has been a net positive contribution from the huge ice sheets of Greenland and Antarctica, but observations of the ice sheets do not yet allow meaningful quantitative estimates of their separate contributions. The ice sheets remain a major source of uncertainty in accounting for past changes in sea level, because there are insufficient data about these ice sheets from 100 years ago.

C.5 Has the climate become more variable and/or extreme?

Many of the impacts of climate change may result from changes in climate variability or extreme weather events. Some reports have already suggested an increase in variability or extremes has taken place in recent decades. Do meteorological records support this?

There are inadequate data to determine whether consistent global changes in climate variability or extremes have occurred over the 20th century. On regional scales there is clear evidence of changes in some extremes and climate variability indicators (e.g., fewer frosts in several widespread areas; an increase in the proportion of rainfall from extreme events over the contiguous states of the USA). Some of these changes have been toward greater variability; some have been toward lower variability.

There have been relatively frequent El Niño-Southern Oscillation warm phase episodes, with only rare excursions into the other extreme of the phenomenon since 1977, as noted in IPCC (1990). This behaviour, and especially the persistent warm phase from 1990 to mid-1995, is unusual in the last 120 years (i.e., since instrumental records began). The relatively low rainfall over the subtropical land areas in the last two decades is related to this behaviour.

D Modelling climate and climate change

Climate models which incorporate, in various degrees of complexity, mathematical descriptions of the atmosphere, ocean, land, biosphere and cryosphere, are important tools for understanding climate and climate change of the past, the present and the future. These models, which use primarily physical laws and physically based empirical relations, are very much more complete than, for example, models based on statistical relationships used in less quantitative disciplines. Detailed projections of future climate change rely heavily on coupled atmosphere-ocean models (see Box 2). How much confidence should we have in predictions from such models?

Box 2:
What tools are used to project future climate and how are they used?

Future climate is projected using climate models. The most highly developed climate models are atmospheric and oceanic general circulation models (GCMs). In many instances GCMs of the atmosphere and oceans, developed as separate models, are combined, to give a *coupled* GCM (termed here a coupled atmosphere-ocean model). These models also include representations of land-surface processes, sea ice related processes and many other complex processes involved in the climate system. GCMs are based upon physical laws that describe the atmospheric and oceanic dynamics and physics and on empirical relationships, and their depiction as mathematical equations. These equations are solved numerically with computers using a three dimensional grid over the globe. For climate, typical resolutions are about 250 km in the horizontal and 1 km in the vertical in atmospheric GCMs, often with higher vertical resolution near the surface and lower resolution in the upper troposphere and stratosphere. Many physical processes, such as those related to clouds, take place on much smaller spatial scales and therefore cannot be properly resolved and modelled explicitly, but their average effects must be included in a simple way by taking advantage of physically based relationships with the larger scale variables (a technique known as parametrization).

Useful weather forecasts can be made using atmospheric GCMs for periods up to about ten days. Such forecasts simulate the evolution of weather systems and describe the associated weather. For simulation and projection of climate, on the other hand, it is the statistics of the system that are of interest rather than the day-to-day evolution of the weather. The statistics include measures of variability as well as mean conditions, and are taken over many weather systems and for several months or more.

When a model is employed for climate projection it is first run for many simulated decades without any changes in external forcing in the system. The quality of the simulation can then be assessed by comparing statistics of the mean climate, the annual cycle and the variability on different time-scales with observations of the current climate. The model is then run with changes in external forcing, for instance with changing greenhouse gas concentrations. The differences between the two climates provide an estimate of the consequent climate change due to changes in that forcing factor. This strategy is intended to simulate changes or perturbations to the system and partially overcomes some imperfections in the models.

Comprehensive coupled atmosphere-ocean models are very complex and take large computer resources to run. To explore all the possible scenarios and the effects of assumptions or approximations in parameters in the model more thoroughly, simpler models are also widely used and are constructed to give results similar to the GCMs when globally averaged. The simplifications may involve coarser resolution, and simplified dynamics and physical processes. An example is the upwelling diffusion-energy balance model. This represents the land and ocean areas in each hemisphere as individual "boxes", with vertical diffusion and upwelling to model heat transport within the ocean.

Early climate experiments, using atmospheric GCMs coupled to a simple representation of the ocean, were aimed at quantifying an equilibrium climate response to a doubling of the concentration of (equivalent) CO_2 in the atmosphere. Such a response portrays the final adjustment of the climate to the changed CO_2 concentration (see Glossary). The range of global warming results is typically between 1.5 and 4.5°C. The temporal evolution and the regional patterns of climate change may depend significantly on the time dependence of the change in forcing. It is important, therefore, to make future projections using plausible evolving scenarios of anthropogenic forcing and coupled atmosphere-ocean models so that the response of the climate to the forcing is properly simulated. These climate simulations are often called "transient experiments" (see Glossary) in contrast to an equilibrium response.

The main uncertainties in model simulations arise from the difficulties in adequately representing clouds and their radiative properties, the coupling between the atmosphere and the ocean, and detailed processes at the land surface.

D.1 The basis for confidence in climate models

As discussed in Section B, changes in the radiatively active trace gases in the atmosphere produce radiative forcing. For equivalent CO_2 concentrations equal to twice the pre-industrial concentration, the positive radiative forcing is about +4 Wm^{-2}. To restore the radiative balance other changes in climate must occur. The initial reaction is for the lower atmosphere (the troposphere) and the Earth’s surface to warm; in the absence of other changes, the warming would be about 1.2°C. However, heating not only changes temperatures, but also alters other aspects of the climate system and various feedbacks are invoked (see Section D.2). The key role of climate models is to quantify these feedbacks and determine the overall climate response. Further, the warming and other climate effects will not be uniform over the Earth's surface; an important role of models is to simulate possible continental and regional scale climate responses.

Climate models include, based on our current understanding, the most important large scale physical processes governing the climate system. Climate models have improved since IPCC (1990), but so too has our understanding of the complexity of the climate system and the recognition of the need to include additional processes.

In order to assess the value of a model for projections of future climate, its simulated climate can be compared with known features of the observed current climate and, to a less satisfactory degree, with the more limited information from significantly different past climate states. It is important to realise that even though a model may have deficiencies, it can still be of value in quantifying the climate response to anthropogenic climate forcing (see also Box 2). Several factors give us some confidence in the ability of climate models to simulate important aspects of anthropogenic climate change in response to anticipated changes in atmospheric composition:

(i) The most successful climate models are able to simulate the important large-scale features of the components of the climate system well, including seasonal, geographical and vertical variations which are a consequence of the variation of forcing and dynamics in space and time. For example, Figure 13 shows the geographical distribution of December to February surface temperature and June to August precipitation simulated by comprehensive coupled atmosphere-ocean models of the type used for climate prediction, compared with observations. The large scale features are reasonably well captured by the models, although at regional scales more discrepancies can be seen. Other seasons are similarly well simulated, indicating the ability of models to reproduce the seasonal cycle in response to changes in solar forcing. The improvement since IPCC (1990) is that this level of accuracy is achieved in models with a fully-interactive ocean as compared to the majority of models that employed simpler schemes used in 1990.

Figure 13
(Colour; Hard copy; full CUP page, landscape)

(ii) Many climate changes are consistently projected by different models in response to greenhouse gases and aerosols and can be explained in terms of physical processes which are known to be operating in the real world, for example, the maximum warming in high northern latitudes in winter (see Section F).

(iii) The models reproduce with reasonable fidelity other less obvious variations in climate due to changes in forcing:

- Some atmospheric models when forced with observed sea surface temperature variations can reproduce with moderate to good skill several regional climate variations, especially in parts of the tropics and sub-tropics. For example, aspects of the large scale interannual atmosphere fluctuations over the tropical Pacific relating to the El Niño-Southern Oscillation phenomenon are captured, as are interannual variations in rainfall in north-east Brazil and to some extent decadal variations in rainfall over the Sahel.
- As discussed in IPCC (1994), stratospheric aerosols resulting from the eruption of Mount Pinatubo in June 1991 gave rise to a short-lived negative global mean radiative forcing of the troposphere which peaked at -3 to -4 Wm^{-2} a few months after the eruption and had virtually disappeared by about the end of 1994. A climate model was used to predict global temperature variations between the time of the eruption and the end of 1994 and the results agreed closely with observations (Figure 14). Such agreement increases confidence in the ability of climate models to respond in a realistic way to transient, planetary-scale radiative forcings of large magnitude.

Figure 14
(B&W; Postscript file; 1 column wide)

- Previous IPCC reports demonstrated the ability of models to simulate some known features of palaeoclimate. Only modest progress has been made in this area, mainly because of the paucity of reliable data for comparison.
- Currently available model simulations of global mean surface temperature trend over the past half century show closer agreement with observations when the simulations include the likely effect of aerosol in addition to greenhouse gases (Figure 15).

Figure 15
(B&W; Postscript file; 1 column wide)

(iv) The model results exhibit "natural" variability on a wide range of time- and space-scales which is broadly comparable to that observed. This "natural" variability arises from the internal processes at work in the climate system and not from changes in external forcing. Variability is a very important aspect of the behaviour of the climate system and has important implications for the detection of climate change (see Section E). The year to year variations of surface air temperature for the current climate are moderately realistic in model simulations at the larger space-scales. For example the smaller variability over the oceans compared with continental interiors is captured. Too low interannual variability of the tropical east and central Pacific Ocean temperatures associated with the El Niño-Southern Oscillation (ENSO) phenomenon is one deficiency. No current coupled atmosphere-ocean model simulates all aspects of ENSO events, but some of the observed interannual variations in the atmosphere associated with these events are captured.

Climate models are calibrated, in part, by introducing adjustments which are empirically determined. The most striking example of these are the systematic adjustments (the so-called flux adjustments) that are used in some models at the atmosphere-ocean interface in order to bring the simulated climate closer to the observed state. These adjustments are used to compensate for model errors, for example inadequacies in the representation of clouds in atmospheric models. Flux adjustments, which can be quite large in some models, are used to ensure that the simulated present day climate is realistic and hence that climate feedback processes operate in the appropriate range in climate change simulations. Many features of the response of models with and without flux adjustments to increasing greenhouse gases are qualitatively similar. The most substantial differences in simulated climate change can generally be traced to deficiencies in the simulation of current climate in the unadjusted models, for example, systematic errors in sea ice. The main unknown regarding the use of adjustments in models is the extent to which they allow important non-linear processes to operate in the models. They have been tested with a good degree of success against known climate variations including the seasonal cycle and the perturbations mentioned above. This provides some confidence in their use for future climate perturbations caused by human activities.

In summary, confidence in climate models has increased since 1990. Primary factors that have served to raise our confidence are model improvements, e.g., the successful incorporation of additional physical processes (such as cloud microphysics and the radiative effects of sulphate aerosols) into global coupled models, and the improvement in such models' simulation of the observed changes in climate over recent decades. Further confidence will be gained as models continue to improve.

D.2 Climate model feedbacks and uncertainties

Warming from radiative forcing will be modified by climate feedbacks which may either amplify (a positive feedback) or reduce (a negative feedback) the initial response. The likely equilibrium response of global surface temperature to a doubling of equivalent carbon dioxide concentration (the "climate sensitivity") was estimated in 1990 to be in the range 1.5 to 4.5 °C, with a "best estimate" of 2.5°C. The range of the estimate arises from uncertainties in the climate models and in their internal feedbacks, particularly those concerning clouds and related processes. No strong reasons have emerged to change these estimates of the climate sensitivity. The present activities regarding incorporation of these feedback processes in models are described below.

Water vapour feedback.

An increase in the temperature of the atmosphere increases its water holding capacity and is expected to be accompanied by an increase in the amount of water vapour. Since water vapour is a powerful greenhouse gas, the increased water vapour would in turn lead to a further enhancement of the greenhouse effect (a positive feedback).

About half of this feedback depends on water vapour in the upper troposphere, whose origin and response to surface temperature increase is not fully understood. Feedback by water vapour in the lower troposphere is unquestionably positive and the preponderance of evidence points to the same conclusion for upper tropospheric water vapour. Feedbacks resulting from changes in the decrease of temperature with height can be comparable to, and partially compensate, the water vapour feedback.

Cloud/radiative feedback

Several processes are involved in cloud/radiative feedback. Clouds can both absorb and reflect solar radiation (which cools the surface) and absorb and emit long-wave radiation (which warms the surface), depending on cloud height, thickness and cloud radiative properties. The radiative properties of clouds depend on the evolution of atmospheric water in its vapour, liquid and ice phases and upon atmospheric aerosols. The processes are complex and, although considerable progress has been made since IPCC (1990) in describing and modelling those cloud processes that are most important for determining radiative and hence temperature changes, their uncertainty represents a significant source of potential error in climate simulation.

This potential error can be estimated by first noting that if clouds and sea ice are kept fixed according to their observed distributions and properties, climate models would all report climate sensitivities in the range of 2 to 3°C. Modellers have shown for various assumptions that physically plausible changes in cloud distribution could either as much as double the warming expected for fixed clouds or, on the other hand, reduce it by up to 1°C. The range in estimated climate sensitivity of 1.5 to 4.5°C is largely dictated by this uncertainty.

Ocean circulation

Oceans play an important role in climate because they carry large amounts of heat from the tropics to the poles. They also store large amounts of heat, carbon and CO_2 and are a major source of water to the atmosphere (through evaporation). Coupling of atmospheric and oceanic GCMs (see Box 2) improves the physical realism of models used for projections of future climate change, particularly the timing and regional distribution of the changes.

Several models show a decrease or only marginal increase of sea surface temperatures in the northern North Atlantic in response to increasing greenhouse gases, related to a slowing down of the thermohaline circulation as the climate warms. This represents a local negative temperature feedback, although changes in cloud cover might be an important factor. The main influence of the oceans on simulations of climate change occurs because of their large heat capacity, which introduces a delay in warming that is not uniform spatially.

Ice and snow albedo feedback

An ice or snow covered surface strongly reflects solar radiation (i.e., it has a high "albedo"). As some ice melts at the warmer surface, less solar radiation is reflected leading to further warming (a positive feedback), but this is complicated by clouds, leads (areas of open water in sea ice) and snow cover.

The realism of simulated sea ice cover varies considerably between models, although sea ice models that include ice dynamics are showing increased accuracy.

Land-surface/atmosphere interactions

Anthropogenic climate changes, e.g., increased temperature, changes in precipitation, changes in net radiative heating and the direct effects of CO_2, will influence the state of the land surface (soil moisture, albedo, roughness, vegetation). In turn, the altered land surface can feed back and alter the overlying atmosphere (precipitation, water vapour, clouds). Changes in the composition and structure of ecosystems can not only alter physical climate, but also the biogeochemical cycles (see Section B). Although land-surface schemes used in current GCMs may be more sophisticated than in IPCC (1990), the disparity between models in their simulation of soil moisture and surface heat and moisture fluxes has not been reduced. Confidence in calculation of regional projections of soil moisture changes in response to greenhouse gas and aerosol forcing remains low.

Changes in vegetation can potentially further modify climate locally and regionally by altering the exchange of water and energy between the land surface and atmosphere. For example, forests spreading into tundra in a warmer world would absorb a greater proportion of solar energy and increase the warming. This feedback would be modified by land-use changes such as deforestation. Coupled atmosphere-ocean models used for climate change studies do not yet include such interactions between climate and vegetation, and such feedbacks may have important effects on regional climate change projection.

E Detection of Climate Change and Attribution of Causes

An important question is: does the instrumental record of temperature change show convincing evidence of a human effect on global climate? With respect to the increase in global mean temperature over the last 100 years, IPCC (1990) concluded that the observed warming was "broadly consistent with predictions of climate models, but it is also of the same magnitude as natural climate variability". The report went on to explain that "the observed increase could be largely due to this natural variability; alternatively this variability and other human factors could have offset a still larger human-induced greenhouse warming".

Since IPCC (1990), considerable progress has been made in the search for an identifiable human-induced effect on climate.

E.1 Better simulations for defining a human-induced climate change "signal"

Experiments with GCMs are now starting to incorporate some of the forcing due to human-induced changes in sulphate aerosols and stratospheric ozone. The inclusion of these additional factors has modified in important ways the picture of how climate might respond to human influences. Furthermore, we now have information on both the timing and spatial patterns of human-induced climate change from a large (>18) number of transient experiments in which coupled atmosphere-ocean models are driven by past and/or projected future time-dependent changes in CO_2 concentration (used as a surrogate to represent the combined effect of CO_2 concentration and other well-mixed greenhouse gases; see "equivalent CO_2" in the Glossary.). Some of these experiments have been repeated with identical forcing but starting from a slightly different initial climate state. Such repetitions help to better define the expected climate response to increasing greenhouse gases and aerosols. However, important uncertainties remain; for example, no model has incorporated the full range of anthropogenic forcing effects.

E.2 Better simulations for estimating natural internal climate variability

In observed data, any "signal" of human effects on climate must be distinguished from the background "noise" of climate fluctuations that are entirely natural in origin. Such natural fluctuations occur on a variety of space- and time-scales, and can be purely internal (due to complex interactions between individual components of the climate system, such as the atmosphere and ocean) or externally-driven by changes in solar variability or the volcanic aerosol loading of the atmosphere. It is difficult to separate a signal from the noise of natural variability in the observations. This is because there are large uncertainties in the evolution and magnitude of both human and natural forcings, and in the characteristics of natural internal variability, which translate to uncertainties in the relative magnitudes of signal and noise.

In the modelled world, however, it is possible to perform multi-century control experiments with no human-induced changes in greenhouse gases, sulphate aerosols or other anthropogenic forcings. Since 1990, a number of such control experiments have been performed with coupled atmosphere-ocean models. These yield important information on the patterns, time-scales, and magnitude of the "internally-generated" component of natural climate variability. This information is crucial for assessing whether observed changes can be plausibly explained by internal climatic fluctuations, but constitutes only one part of the "total" natural variability of climate (since such control runs do not include changes in solar output or volcanic aerosols). Uncertainties still remain in estimates of both internal and total natural climate variability, particularly on the decadal-to-century time-scales.

E.3 Studies of global mean change

Most studies that have attempted to detect an anthropogenic effect on climate have used changes in global mean, annually-averaged temperature only. These investigations compared observed changes over the past 10-100 years with estimates of internal or total natural variability noise derived from palaeodata, climate models, or statistical models fitted to observations. Most but not all of these studies show that the observed change in global mean, annually-averaged temperature over the last century is unlikely to be due entirely to natural fluctuations of the climate system.

These global mean results cannot establish a clear cause and effect link between observed changes in atmospheric greenhouse gas concentrations and changes in the Earth's surface temperature. This is the attribution issue. Attribution is difficult using global mean changes only because of uncertainties in the histories and magnitudes of natural and human-induced forcings: there are many possible combinations of these forcings that could yield the same curve of observed global mean temperature change. Some combinations are more plausible than others, but relatively few data exist to constrain the range of possible solutions. Nevertheless, model-based estimates of global temperature increase over the last 130 years are more easily reconciled with observations when the likely cooling effect of sulphate aerosols is taken into account, and provide qualitative support for an estimated range of climate sensitivity consistent with that given in IPCC (1990) (Figure 16).

Figure 16
(B&W; Postscript file; 1 column wide)

E.4 Studies of patterns of change

To better address the attribution problem, a number of recent studies have compared observations with model-predicted *patterns* of temperature-change in response to anthropogenic forcing. The argument underlying pattern-based approaches is that different forcing mechanisms ("causes") may have different patterns of response ("effects"), particularly if one considers the full three- or even four-dimensional structure of the response, e.g., temperature change as a function of latitude, longitude, height and time. Thus a good match between modelled and observed multi-dimensional patterns of climate change would be difficult to achieve for "causes" other than those actually used in the model experiment.

Several studies have compared observed patterns of temperature change with model patterns of change from simulations with changes in both greenhouse gases and anthropogenic sulphate aerosols. These comparisons have been made at the Earth's surface and in vertical sections through the atmosphere. While there are concerns regarding the relatively simple treatment of aerosol effects in model experiments, and the neglect of other potentially significant contributions to the radiative forcing, all such pattern comparison studies show significant correspondence between the observations and model predictions (an example is shown in Figure 17). Much of the model-observed correspondence in these experiments occurs at the largest spatial scales - for example, temperature differences between hemispheres, land and ocean, or the troposphere and stratosphere. Model predictions are more reliable at these spatial scales than at the regional scale. Increasing confidence in the identification of a human-induced effect on climate comes primarily from such pattern-based work. For those seasons during which aerosol effects should be most pronounced the pattern correspondence is generally higher than that achieved if model predictions are based on changes in greenhouse gases alone (Figure 17).

Figure 17
(Colour; Postscript file; 2 columns wide)

As in the global mean studies, pattern-oriented detection work relies on model estimates of internal natural variability as the primary yardstick for evaluating whether observed changes in temperature patterns could be due to natural causes. Concerns remain regarding the reliability of this yardstick.

E.5 Qualitative consistency

In addition to quantitative studies, there are broad areas of qualitative agreement between observations and those model predictions that either include aerosol effects or do not depend critically on their inclusion. As in the quantitative studies, one must be cautious in assessing consistency because the expected climate change signal due to human activities is still uncertain, and has changed as our ability to model the climate system has improved. In addition to the surface warming, the model and observed commonalities in which we have most confidence include stratospheric cooling, reduction in diurnal temperature range, sea level rise, high latitude precipitation increases and water vapour and evaporation increase over tropical oceans.

E.6 Overall assessment of the detection and attribution issues

In summary, the most important results related to the issues of detection and attribution are:

- The limited available evidence from proxy climate indicators suggests that the 20th century global mean temperature is at least as warm as any other century since at least 1400 AD. Data prior to 1400 are too sparse to allow the reliable estimation of global mean temperature (see Section C.2).
- Assessments of the statistical significance of the observed global mean temperature trend over the last century have used a variety of new estimates of natural internal and externally-forced variability. These are derived from instrumental data, palaeodata, simple and complex climate models, and statistical models fitted to observations. Most of these studies have detected a significant change and show that the observed warming trend is unlikely to be entirely natural in origin.

- More convincing recent evidence for the attribution of a human effect on climate is emerging from pattern-based studies, in which the modelled climate response to combined forcing by greenhouse gases and anthropogenic sulphate aerosols is compared with observed geographical, seasonal and vertical patterns of atmospheric temperature change. These studies show that such pattern correspondences increase with time, as one would expect as an anthropogenic signal increases in strength. Furthermore, the probability is very low that these correspondences could occur by chance as a result of natural internal variability only. The vertical patterns of change are also inconsistent with those expected for solar and volcanic forcing.
- Our ability to quantify the human influence on global climate is currently limited because the expected signal is still emerging from the noise of natural variability, and because there are uncertainties in key factors. These include the magnitude and patterns of long-term natural variability and the time-evolving pattern of forcing by, and response to, changes in concentrations of greenhouse gases and aerosols, and land-surface changes. Nevertheless, the balance of evidence suggests that there is a discernible human influence on global climate.

F The prospects for future climate change

F.1 Forcing scenarios

Projections of future anthropogenic climate change depend, amongst other things, on the scenario used to force the climate model. The IS92 emission scenarios are used here for projections of changes in global mean temperature and sea level. The IS92 scenarios include emissions of both greenhouse gases and aerosol precursors (see Section B.9.1) and for the first time both factors have been taken into account in the global mean temperature and sea level projections (Section F.2).

In many coupled model experiments the forcing scenario is simplified by summing the radiative forcings of all the trace gases (CO_2, CH_4, O_3, etc.) and treating the total forcing as if it came from an "equivalent" concentration of CO_2. The rate of increase of "equivalent CO_2" in these experiments is often assumed to be a constant +1%/yr (compounded). For comparison the IS92a Scenario, neglecting the effect of aerosols, is equivalent to a compounded rate of increase varying from 0.77 to 0.84%/yr during the 21st century.

The projections of global mean temperature and sea level changes do not come directly from coupled atmosphere-ocean models. Though these are the most sophisticated tools available for making projections of future climate change they are computationally expensive, making it unfeasible to produce results based on a large number of emission scenarios. In order to assess global temperature and sea level projections for the full range of IS92 emission scenarios, simple upwelling diffusion-energy balance models (see Box 2) can be employed to interpolate and extrapolate the coupled model results. These models, used for similar tasks in IPCC (1990) and IPCC (1992), are calibrated to give the same globally averaged temperature response as the coupled atmosphere-ocean models.

The climate simulations here are called *projections* instead of *predictions* to emphasise that they do not represent attempts to forecast the most likely (or "best estimate") evolution of climate in the future. The projections are aimed at estimating and understanding responses of the climate system to possible forcing scenarios.

F.2 Projections of climate change

F.2.1 Global mean temperature response to IS92 emission scenarios

Using the IS92 emission scenarios, which include emissions of both greenhouse gases and aerosol precursors (Section B.9.1) projected global mean temperature changes relative to 1990 were calculated for the 21st century. Temperature projections assuming the "best estimate" value of climate sensitivity, 2.5°C, (see Section D.2) are shown for the full set of IS92 scenarios in Figure 18. For IS92a the temperature increase by 2100 is about 2°C. Taking account of the range in the estimate of climate sensitivity (1.5 to 4.5°C) and the full set of IS92 emission scenarios, the models project an increase in global mean temperature of between 0.9 and 3.5°C (Figure 19). In all cases the average rate of warming would probably be greater than any seen in the last 10,000 years, but the actual annual to decadal changes would include considerable natural variability. Because of the thermal inertia of the oceans, global mean temperature would continue to increase beyond 2100 even if concentrations of greenhouse gases were stabilised at that time. Only 50-90% of the eventual temperature changes are realised at the time of greenhouse gas stabilisation. All scenarios show substantial climate warming, even when the negative aerosol radiative forcing is accounted for. Although CO_2 is the most important anthropogenic greenhouse gas, the other greenhouse gases contribute significantly (about 30%) to the projected global warming.

Figure 18
(B&W; Postscript file; 1 column wide)
Figure 19
(B&W; Postscript file; 1 column wide)

To allow closer comparison with the projections presented in IPCC (1990) and IPCC (1992) and to illustrate the sensitivity of future global temperature to changes in aerosol concentrations, the same series of calculations were performed neglecting future aerosol changes, i.e. aerosol concentrations were held constant at 1990 levels. These lead to higher projections of temperature change. Taking account of the range in the estimate of climate sensitivity and the full set of IS92 emission scenarios, the models project an increase in global mean temperature of between 0.8 and 4.5°C. For IS92a, assuming the "best estimate" of climate sensitivity, the temperature increase by 2100 is about 2.4°C. For comparison, the corresponding temperature increase for IS92a presented in IPCC (1992) was 2.8°C. The projections in IPCC (1990) were based on an earlier set of emission scenarios, the "best estimate" for the increase in global temperature by 2100 (relative to 1990) was 3.3°C.

F.2.2 Global mean sea level response to IS92 emission scenarios

Using the IS92 emission scenarios, including greenhouse gas and aerosol precursors, projected global mean sea level increases relative to 1990 were calculated for the 21st century. Sea level projections assuming the "best estimate" values for climate sensitivity and ice melt are shown for the full set of IS92 scenarios in Figure 20. For IS92a, the sea level rise by 2100 is about 49 cm. For comparison, the "best estimate" of global sea level rise by 2100 given in IPCC (1990) was about 65 cm. Also taking account of the ranges in the estimate of climate sensitivity and ice melt parameters, and the full set of IS92 emission scenarios, the models project an increase in global mean sea level of between 15 and 95 cm (Figure 21). During the first half of the next century, the choice of emission scenario has relatively little effect on the projected sea level rise due to the large thermal inertia of the ocean-ice-atmosphere climate system, but increasingly larger effects in the latter part of the next century. In addition, because of the thermal inertia of the oceans, sea level would continue to rise for many centuries beyond 2100 even if concentrations of greenhouse gases were stabilised at that time. The projected rise in sea level is primarily due to thermal expansion as the ocean waters warm, but also due to increased melting of glaciers.

Figure 20
(B&W; Postscript file; 1 column wide)
Figure 21
(B&W; Postscript file; 1 column wide)

In these projections, the combined contributions of the Greenland and Antarctic ice sheets are projected to be relatively minor over the next century. However, the possibility of large changes in the volumes of these ice sheets (and, consequently, in sea level) cannot be ruled out, although the likelihood is considered to be low.

Changes in future sea level will not occur uniformly around the globe. Recent coupled atmosphere-ocean model experiments suggest that the regional responses could differ substantially, owing to regional differences in heating and circulation changes. In addition, geological and geophysical processes cause vertical land movements and thus affect relative sea levels on local and regional scales.

Tides, waves and storm surges could be affected by regional climate changes, but future projections are, at present, highly uncertain.

F.2.3 Temperature and sea level projections compared with IPCC (1990)

The global average temperature and sea level projections presented here for 1990 to 2100, both excluding and including changing aerosol emissions, are lower than the corresponding projections presented in IPCC (1990). Taking into account the negative radiative forcing of aerosols reduces projections of temperature and sea level rise. Those projections which *exclude* the effect of changing aerosol emissions are lower than IPCC (1990) for a number of reasons, mainly:

- The revised (IS92) emission scenarios have been used for all greenhouse gases. This is particularly important for CO_2 and CFCs.
- Revised treatment of the carbon cycle. The carbon cycle model used to calculate future temperature and sea level rise in IPCC (1990) and IPCC (1992) did not incorporate the effect of

carbon uptake through CO_2 fertilisation, resulting in higher future CO_2 concentrations for given emissions in IPCC (1990).

- The inclusion of aerosol effects in the pre-1990 radiative forcing history. The estimated historical changes of radiative forcing up to 1990, used in this report for global mean temperature and sea level projections, includes a component due to aerosols. This particularly affects projections of sea level rise, which are strongly influenced by the history of radiative forcing over the last century.
- Revised (and more realistic) parameters in the simple upwelling diffusion-energy balance climate model.
- The inclusion in the model of spatial variations in the climate sensitivity and the effect of changing strength of the thermohaline circulation, to accord with coupled atmosphere-ocean general circulation models.
- The use of improved models for the ice melt component of sea level rise.

F.3 Spatial patterns of projected climate change

Although in *global mean* terms, the effect of including aerosols is to reduce the projected warming (see Section F.2), it can be misleading to consider only the global mean surface temperature, which does not give an effective indication of climate change at smaller spatial scales.

Because aerosols are short-lived, they are unevenly distributed across the globe, being concentrated near regions where they are emitted. As a result, the spatial pattern of aerosol forcing is very different to that produced by the long-lived well-mixed greenhouse gases and, when considering patterns of climate change, their cooling effect is not a simple offset to the warming effect of greenhouse gases, as might be implied from the global mean results. Aerosols are likely to have a significant effect on future regional climate change.

Confidence is higher in hemispheric to continental scale projections of climate change (Section F.3.1) than at regional scales (Section F.3.2), where confidence remains low.

F.3.1 Continental scale patterns

In IPCC (1990), estimates of the patterns of future climate change were presented, the most robust of which related to continental and larger spatial scales. The results were based on GCM experiments which included the effect of greenhouse gases, but did not take into account the effects of aerosols.

The following provides some details of the changes on continental scales in experiments with greenhouse gases alone (generally represented by a 1%/yr increase in CO_2) and increases in greenhouse gas and aerosol concentrations (using aerosol concentration derived from Scenario IS92a). It is important to realise that, in contrast to the many model results with CO_2 alone, there are only two recent coupled atmosphere-ocean model simulations that include the effects of both aerosols and CO_2, neither of which have yet been thoroughly analysed. We have concentrated on those changes which show most consistency between models, and for which plausible physical mechanisms have been identified.

Temperature and Precipitation

All model simulations, whether they are forced with increased concentrations of greenhouse gases and aerosols, or with increased greenhouse gas concentrations alone, show the following features:

- generally greater surface warming of the land than of the oceans in winter, as in equilibrium simulations (Figures 22 and 23);
- a minimum warming around Antarctica and in the northern North Atlantic which is associated with deep oceanic mixing in those areas;
- maximum annual mean warming in high northern latitudes in late autumn and winter associated with reduced sea ice and snow cover;
- little warming over the Arctic in summer;
- little seasonal variation of the warming in low latitudes or over the southern circumpolar ocean;
- a reduction in diurnal temperature range over land in most seasons and most regions;
- an enhanced global mean hydrological cycle;
- increased precipitation in high latitudes in winter.

Figure 22
(Colour; Postscript file; 2columns wide)
Figure 23
(Colour; Postscript file; 2 columns wide)

Including the effects of aerosols in simulations of future climate leads to a somewhat reduced surface warming, mainly in middle latitudes of the Northern Hemisphere. The maximum winter warming in high northern latitudes is less extensive (compare Figures 22 and 23)

However, adding the cooling effect of aerosols is not a simple offset to the warming effect of greenhouse gases, but significantly affects some of the continental scale patterns of climate change. This is most noticeable in summer where the cooling due to aerosols tends to weaken monsoon circulations. For example, when the effects of both greenhouse gases and aerosols are included, Asian summer monsoon rainfall decreases, whereas in earlier simulations with only the effect of greenhouse gases represented, Asian summer monsoon rainfall increased. Conversely, the addition of aerosol effects leads to an increase in precipitation over southern Europe, whereas decreases are found in simulations with greenhouse gases only. These changes will be sensitive to the aerosol scenario used, and the details of the parametrization of the radiative effects of aerosol. Other forcings, including that due to increases in tropospheric ozone, soot and the indirect effect of sulphate aerosols have been neglected and could influence these results. In general, regional projections are also sensitive to model resolution and are affected by large natural variability. Hence confidence in regional projections remains low.

With increases in CO_2 only, two coupled atmosphere-ocean models show a pattern of SST (sea surface temperature) change, precipitation change and anomalies in wind and ocean currents that resemble the warm phase of ENSO, as well as the observed decadal time-scale SST anomalies of the

1980s and early 1990s. This is characterised by a reduction of the east-west SST gradient in the tropical Pacific though the magnitude of this effect varies among models.

Soil moisture

Although there is less confidence in simulated changes in soil moisture than in those of temperature, some of the results concerning soil moisture are dictated more by changes in precipitation and evaporation than by the detailed response of the surface scheme of the climate model. All model simulations, whether they are forced with increased concentrations of greenhouse gases and aerosols, or with increased greenhouse gas concentrations alone, produce predominantly increased soil moisture in high northern latitudes in winter. Over the northern continents in summer, the changes in soil moisture are sensitive to the inclusion of aerosol effects.

Ocean circulation

In response to increasing greenhouse gases, most models show a decrease in the strength of the northern North Atlantic oceanic circulation further reducing the strength of the warming around the North Atlantic. The increase in precipitation in high latitudes decreases the surface salinity, inhibiting the sinking of water at high latitude, which drives this circulation.

F.3.2 Regional scale patterns

Estimation of the potential impacts of climate change on human infrastructure and natural ecosystems requires projections of future climate changes at the regional scale, rather than as global or continental means.

Since IPCC (1990), a greater appreciation has been developed of the uncertainties in making projections at the regional scale. There are several difficulties:

- The global climate models used for future projections are run at fairly coarse resolution and do not adequately depict many geographic features (such as coastlines, lakes and mountains), surface vegetation, and the interactions between the atmosphere with the surface which become more important on regional scales. Considerable spread exists among model projections on the regional scale even when climate model experiments are driven by the same future radiative forcing scenario.
- There is much more natural variation in local climate than in climate averaged over continental or larger scales. This variation arises from locally-generated variability from storms, interactions between the atmosphere and the oceans (such as ENSO), and from variations in soil moisture, sea ice, and other components of the climate system. Series or ensembles of model predictions started from different initial conditions allow both the mean climate and the superimposed variability to be determined.
- Because of their uneven spatial distribution, human induced tropospheric aerosols are likely to greatly influence future regional climate change. At present, however, there are very few projections of climate change with coupled atmosphere-ocean models (the type of model that

gives more reliable information on the regional scale) which include the radiative effects of aerosols. Those that have been run include a very simplified representation of aerosol effects.

- Land-use changes are also believed to have a significant impact on temperature and precipitation changes, especially in the tropics and subtropics. Climate model experiments have shown the likelihood of substantial local climate change associated with deforestation in the Amazon, or desertification in the Sahel. Changes in land-use on small scales which cannot be foreseen are expected to continue to influence regional climate.

Because of these problems, no information on future regional climate change is presented here. However, this situation is expected to improve in the future as a result of:

- more coupled atmosphere-ocean model experiments with aerosol effects included;
- improvements in models, both from increased resolution and improved representation of small-scale processes;
- More refined scenarios for aerosols and other forcings.

F.3.3 Changes in Variability and extremes

Small changes in the mean climate or climate variability can produce relatively large changes in the frequency of extreme events (defined as events where a certain threshold is surpassed); a small change in the variability has a stronger effect than a similar change in the mean.

Temperature

A general warming tends to lead to an increase in extremely high temperature events and a decrease in extremely low temperatures (e.g., frost days).

Hydrology

Many models suggest an increase in the probability of intense precipitation with increased greenhouse gas concentrations. In some areas a number of simulations show there is also an increase in the probability of dry days and the length of dry spells (consecutive days without precipitation). Where mean precipitation decreases, the likelihood of drought increases. New results reinforce the view that variability associated with the enhanced hydrological cycle translates into prospects for more severe droughts and/or floods in some places and less severe droughts and/or floods in other places.

Mid-latitude storms

In the few analyses available, there is little agreement between models on the changes in storminess that might occur in a warmer world. Conclusions regarding extreme storm events are obviously even more uncertain.

Hurricanes/Tropical cyclones

The formation of tropical cyclones depends not only on sea surface temperature (SST), but also on a number of atmospheric factors. Although some models now represent tropical storms with some realism for present day climate, the state of the science does not allow assessment of future changes.

El Niño-Southern Oscillation

Several global coupled models indicate that the ENSO-like SST variability they simulate continues with increased CO_2. Associated with the mean increase of tropical SSTs as a result of increased greenhouse gas concentrations, there could be enhanced precipitation variability associated with ENSO events in the increased CO_2 climate, especially over the tropical continents.

F.4 Effects of stabilising greenhouse gas concentrations

Possible global temperature and sea level response to the scenarios for stabilising concentrations discussed in Section B.9.2 were calculated with the same upwelling diffusion-energy balance model used for the results in Sections F.2.1 and F.2.2.

For each of the constructed stabilising emissions scenarios, the climate system shows considerable warming during the 21st century. Figure 24 shows temperature increases for the cases which stabilise at concentrations of 650 and 450 ppmv for different climate sensitivities. Stabilisation of the concentration does not lead to an immediate stabilisation of the global mean temperature. The global mean temperature is seen to continue rising for hundreds of years after the concentrations have stabilised in Figure 24 due to long time-scales in the ocean.

Figure 24
(B&W; Postscript file; 1 column wide)

As shown in Figure 25 , the long-term sea level rise "commitment" is even more pronounced. Sea level continues to rise, at only a slowly declining rate, for many centuries after greenhouse gas concentrations and temperatures have stabilised.

Figure 25
(B&W; Postscript file; 1 column wide)

F.5 The possibility of surprises

Unexpected external influences, such as volcanic eruptions, can lead to unexpected and relatively sudden shifts in the climatic state. Also, as the response of the climate system to various forcings can be non-linear, its response to gradual forcing changes may be quite irregular. Abrupt and significant changes in the atmospheric circulation involving the North Pacific which began about 1976 were described in IPCC (1990). A related example is the apparent fluctuation in the recent behaviour of ENSO, with warm conditions prevailing since 1989, a pattern which has been unusual compared to previous ENSO behaviour.

Another example is the possibility that the West Antarctic ice sheet might "surge", causing a rapid rise in sea level. The current lack of knowledge regarding the specific circumstances under which this might occur, either in total or in part, limits the ability to quantify this risk. Nonetheless, the likelihood of a major sea level rise by the year 2100 due to the collapse of the West Antarctic ice sheet is considered low.

In the oceans the meridional overturning might weaken in a future climate. This overturning (the thermohaline circulation) is driven in part by deep convection in the northern North Atlantic Ocean and keeps the northern North Atlantic Ocean several degrees warmer than it would otherwise be. Both the study of palaeoclimate from sediment records and ice cores and modelling studies with coupled climate models and ocean GCMs can be interpreted to suggest that the ocean circulation has been very different in the past. Both in these observations and in the ocean models, transitions between different types of circulation seem to occur on a time-scale of a few decades, so relatively sudden changes in the regional (North Atlantic, Western Europe) climate could occur, presumably mainly in response to precipitation and runoff changes which alter the salinity, and thus the density, of the upper layers of the North Atlantic. Whether or not such a sudden change can actually be realised in response to global warming and how strong a perturbation is required to cause a transition between types of circulation are still the subject of much debate.

In terrestrial ecological systems, there are thresholds in the sustained temperature and water availability at which one biological population is replaced by another. Some replacement, e.g., in tree species, is slow while some, e.g., in micro-organisms is rapid. Minimum temperatures exist for the survival of organisms in winter, and the populations of such organisms may move polewards as the climate and especially night-time temperatures warm. If the transitions are not orderly, sudden shifts in ecosystem functioning will occur. These may have impacts of direct human relevance (as discussed in IPCC WGII (1995)) but may also have surprising impacts on climate via effects on albedo, aerosol forcing, the hydrological cycle, evapotranspiration, CO_2 release and methane cycling, for example (see Sections B.1 and D.2).

G Advancing our understanding

An important long-term goal is the accurate projection of regional climate change, so that potential impacts can be adequately assessed. Progress towards this objective depends on determining the likely global magnitude and rate of human-induced climate change, including sea level change, as well as the regional expressions of these quantities. The detection and attribution of human-induced climate change is also most important. To achieve these objectives requires systematic and sustained global observations of relevant variables, as well as requiring the effective co-operation and participation of many nations. The most urgent scientific problems requiring attention concern:

(i) the *rate* and *magnitude* of climate change and sea level rise:

- the factors controlling the distribution of clouds and their radiative characteristics;
- the hydrological cycle, including precipitation, evaporation and runoff;
- the distribution and time evolution of ozone and aerosols and their radiative characteristics;

- the response of terrestrial and marine systems to climate change and their positive and negative feedbacks;
- the response of ice sheets and glaciers to climate;
- the influence of human activities on emissions;
- the coupling between the atmosphere and ocean, and ocean circulation;
- the factors controlling the atmospheric concentrations of carbon dioxide and other greenhouse gases;

(ii) *detection* and *attribution* of climate change:

- systematic observations of key variables, and development of model diagnostics relating to climate change;
- relevant proxy data to construct and test palaeoclimatic time-series to describe natural variability of the climate system;

(iii) *regional patterns* of climate change:

- land-surface processes and their link to atmospheric processes;
- coupling of scales between global climate models and regional and smaller scale models;
- simulations with higher resolution climate models.

The research activities for each objective are strongly interconnected. Such research is and needs to be conducted by individual investigators in a variety of institutions, as well as by co-ordinated international efforts which pool national resources and talents in order to more efficiently engage in large-scale integrated field and modelling programmes to advance our understanding.

Glossary

Term	*Definition*
Aerosols	Airborne particles. The term has also come to be associated, erroneously, with the propellant used in "aerosol sprays".
Climate change (FCCC usage)	A change of climate which is attributed directly or indirectly to human activity that alters the composition of the global atmosphere and which is in addition to natural climate variability observed over comparable time periods.
Climate change (IPCC usage)	Climate change as referred to in the observational record of climate occurs because of internal changes within the climate system or in the interaction between its components, or because of changes in external forcing either for natural reasons or because of human activities. It is generally not possible clearly to make attribution between these causes. Projections of future climate change reported by IPCC generally consider only the influence on climate of anthropogenic increases in greenhouse gases and other human-related factors.
Climate sensitivity	In IPCC reports, climate sensitivity usually refers to the long-term (equilibrium) change in global mean surface temperature following a doubling of atmospheric CO_2 (or equivalent CO_2) concentration. More generally, it refers to the equilibrium change in surface air temperature following a unit change in radiative forcing (°C/Wm^{-2}).
Diurnal temperature range	The difference between maximum and minimum temperature over a period of 24 hours
Equilibrium climate experiment	An experiment where a step change is applied to the forcing of a climate model and the model is then allowed to reach a new equilibrium. Such experiments provide information on the difference between the initial and final states of the model, but not on the time-dependent response.

Equivalent CO_2	The concentration of CO_2 that would cause the same amount of radiative forcing as the given mixture of CO_2 and other greenhouse gases.
Evapotranspiration	The combined process of evaporation from the Earth's surface and transpiration from vegetation.
Greenhouse gas	A gas that absorbs radiation at specific wavelengths within the spectrum of radiation (infrared radiation) emitted by the Earth's surface and by clouds. The gas in turn emits infrared radiation from a level where the temperature is colder than the surface. The net effect is a local trapping of part of the absorbed energy and a tendency to warm the planetary surface. Water vapour (H_2O), carbon dioxide (CO_2), nitrous oxide (N_2O), methane (CH_4) and ozone (O_3) are the primary greenhouse gases in the Earth's atmosphere.
Ice-cap	A dome-shaped glacier usually covering a highland near a water divide.
Ice sheet	A glacier more than 50,000 km^2 in area forming a continuous cover over a land surface or resting on a continental shelf.
Radiative forcing	A simple measure of the importance of a potential climate change mechanism. Radiative forcing is the perturbation to the energy balance of the Earth-atmosphere system (in Wm^{-2}) following, for example, a change in the concentrations of carbon dioxide or a change in the output of the Sun; the climate system responds to the radiative forcing so as to re-establish the energy balance. A positive radiative forcing tends to warm the surface and a negative radiative forcing tends to cool the surface. The radiative forcing is normally quoted as a global and annual mean value. A more precise definition of radiative forcing, as used in IPCC reports, is the perturbation of the energy balance of the surface-troposphere system, after allowing for the stratosphere to re-adjust to a state of global mean radiative equilibrium (see Chapter 4 of IPCC (1994)). Sometimes called "climate forcing".
Spatial scales	continental 10 - 100 million square kilometres (km^2) regional 100 thousand - 10 million km^2 local less than 100 thousand km^2

Soil moisture	Water stored in or at the continental surface and available for evaporation. In IPCC (1990) a single store (or "bucket") was commonly used in climate models. Today's models which incorporate canopy and soil processes view soil moisture as the amount held in excess of plant "wilting point".
Stratosphere	The highly stratified and stable region of the atmosphere above the troposphere (*qv*.) extending from about 10 km to about 50 km.
Thermohaline circulation	Large scale density-driven circulation in the oceans, driven by differences in temperature and salinity.
Transient climate experiment	An analysis of the time-dependent response of a climate model to a time-varying change of forcing.
Troposphere	The lowest part of the atmosphere from the surface to about 10 km in altitude in mid-latitudes (ranging from 9 km in high latitudes to 16 km in the tropics on average) where clouds and "weather" phenomena occur. The troposphere is defined as the region where temperatures generally decrease with height.

References

IPCC , (Intergovernmental Panel on Climate Change) 1990: *Climate Change: The IPCC Scientific Assessment*, J.T. Houghton, G.J. Jenkins and J.J. Ephraums (eds). Cambridge University Press, Cambridge, UK, 365 pp.

IPCC, 1992: *Climate Change 1992: The Supplementary Report to the IPCC Scientific Assessment*, J.T. Houghton, B.A. Callander and S.K. Varney (eds). Cambridge University Press, Cambridge, UK, 198 pp.

IPCC, 1994: *Climate Change 1994: Radiative Forcing of Climate Change and an Evaluation of the IPCC 1S92 Emission Scenarios,* J.T. Houghton, L.G. Meira Filho, J. Bruce, Hoesung Lee, B.A. Callander, E. Haites, N. Harris and K. Maskell (eds). Cambridge University Press, Cambridge, UK, 339 pp.

IPCC WGII, 1995: *Climate Change 1995-Impacts, Adaptations and Mitigations of Climate Change: Scientific-Technical Analyses* , R.T. Watson, M.C. Zinyowera and R.H. Moss (eds). Cambridge University Press, New York, USA.

Figure Captions

Figure 1: (a) CO_2 concentrations over the past 1000 years from ice core records (D47, D57, Siple and South Pole) and (since 1958) from Mauna Loa, Hawaii, measurement site. All ice core measurements were taken in Antarctica. The smooth curve is based on a hundred year running mean. The rapid increase in CO_2 concentration since the onset of industrialisation is evident and has followed closely the increase in CO_2 emissions from fossil fuels (see inset of period from 1850 onwards). (b) Growth rate of CO_2 concentration since 1958 in ppmv/yr at Mauna Loa. The smooth curve shows the same data but filtered to suppress variations on time-scales less than approximately 10 years.

Figure 2: Estimates of the globally and annually averaged anthropogenic radiative forcing (in Wm^{-2}) due to changes in concentrations of greenhouse gases and aerosols from pre-industrial times to the present (1992) and to natural changes in solar output from 1850 to the present. The height of the rectangular bar indicates a mid-range estimate of the forcing whilst the error bars show an estimate of the uncertainty range, based largely on the spread of published values; the "confidence level" indicates the author's confidence that the actual forcing lies within this error bar . The contributions of individual gases to the direct greenhouse forcing is indicated on the first bar. The indirect greenhouse forcings associated with the depletion of stratospheric ozone and the increased concentration of tropospheric ozone are shown in the second and third bar respectively. The direct contributions of individual tropospheric aerosol component are grouped into the next set of three bars. The indirect aerosol effect, arising from the induced change in cloud properties, is shown next; quantitative understanding of this process is very limited at present and hence no bar representing a mid-range estimate is shown. The final bar shows the estimate of the changes in radiative forcing due to variations in solar output. The forcing associated with stratospheric aerosols resulting from volcanic eruptions is not shown, as it is very variable over this time period. Note that there are substantial differences in the geographical distribution of the forcing due to the well-mixed greenhouse gases (mainly CO_2, N_2O, CH_4 and the halocarbons) and that due to ozone and aerosols, which could lead to significant differences in their respective global and regional climate responses. For this reason, the negative radiative forcing due to aerosols should not necessarily be regarded as an offset against the greenhouse gas forcing.

Figure 3: Global methane concentrations (ppbv) for 1983 to 1994. Concentrations observed at Mould Bay, Canada are also shown.

Figure 4: Global CFC-11 concentrations (pptv) for 1978 to 1994. As one of the ozone-depleting gases, the emissions of CFC-11 are controlled under the Montreal Protocol and its Adjustments and Amendments. Observations at some individual measurement sites are also shown.

Figure 5: (a) Total anthropogenic CO_2 emissions under the IS92 emission scenarios and (b) the resulting atmospheric CO_2 concentrations calculated using the "Bern" carbon cycle model and the carbon budget for the 1980s shown in Table 2.

Figure 6: (a) Total globally and annually averaged historical radiative forcing from 1765 to 1990 due to changes in greenhouse gas concentrations and tropospheric aerosol emissions and projected radiative forcing values to 2100 derived from the IS92 emissions scenarios. (b) Radiative forcing components resulting from the IS92a emission scenario for 1990 to 2100. The "Total non-CO_2 trace gases" curve includes the radiative forcing from methane (including methane related increases in stratospheric water vapour), nitrous oxide, tropospheric ozone and the halocarbons (including the negative forcing effect of stratospheric ozone depletion). Halocarbon emissions have been modified to take account of the Montreal Protocol and its Adjustments and Amendments. The three aerosol components are: direct sulphate, indirect sulphate and direct biomass burning. (c) Non-CO_2 trace gas radiative forcing components. "Cl/Br-direct" is the direct radiative forcing resulting from the chlorine and bromine containing halocarbons; emissions are assumed to be controlled under the Montreal Protocol and its Adjustments and Amendments. The indirect forcing from these compounds (through stratospheric ozone depletion) is shown separately (Strat.O_3). All other emissions follow the IS92a Scenario. The tropospheric ozone forcing (Trop.O_3) takes account of concentration changes due only to the indirect effect of methane.

Figure 7: (a) CO_2 concentration profiles leading to stabilisation at 450, 550, 650 and 750 ppmv following the pathways defined in IPCC (1994) (solid curves) and for pathways that allow emissions to follow IS92a until at least 2000 (dashed curves). A single profile that stabilises at a CO_2 concentration of 1000 ppmv and follows IS92a emissions until at least 2000 has also been defined. (b) CO_2 emissions leading to stabilisation at concentrations of 450, 550, 650, 750 and 1000 ppmv following the profiles shown in (a). Current anthropogenic CO_2 emissions and those for IS92a are shown for comparison. The calculations use the "Bern" carbon cycle model and the carbon budget for the 1980s shown in Table 2.

Figure 8: Combined land-surface air and sea surface temperatures (°C) 1861 to 1994, relative to 1961 to 1990. The solid curve represents smoothing of the annual values shown by the bars to suppress sub-decadal time-scale variations. The dashed smoothed curve is the corresponding result from IPCC (1992).

Figure 9: Change (from 1955-74 to 1975-94) of annual land-surface air temperature and sea surface temperature.

Figure 10: Decadal summer (June to August) temperature index for the Northern Hemisphere (to 1970-1979) based on 16 proxy records (tree-rings, ice cores, documentary records) from North

America, Europe and East Asia. The thin line is a smoothing of the same data. Anomalies are relative to 1961 to 1990.

Figure 11: Changes in land-surface precipitation averaged over regions between 55°S and 85°N. Annual precipitation departures from the 1961-90 period are depicted by the hollow bars. The continuous curve is a smoothing of the same data.

Figure 12: Summary of observed climatic trends during the instrumental period of record.

Figure 13: The geographical distribution of December to February surface temperature (a) and June to August precipitation (c) compared to that simulated by comprehensive coupled models of the type used for climate projection (b) and (d).

Figure 14: Predicted and observed changes in global land and ocean surface air temperature after the eruption of Mt. Pinatubo. Lines represent changes of three month running mean temperature from April to June 1991 until March to May 1995. The two model lines represent predictions starting from different initial atmospheric conditions.

Figure 15: Simulated global annual mean warming from 1860 to 1990 allowing for increases in greenhouse gases only (dashed curve) and greenhouse gases and sulphate aerosols (solid curve), compared with observed changes over the same period.

Figure 16 Observed changes in global mean temperature over 1861 to 1994 compared with those simulated using an upwelling diffusion-energy balance climate model. The model was run first with forcing due to greenhouse gases alone (a) and then with greenhouse gases and aerosols (b).

Figure 17 Annual mean near-surface air temperature changes (^{o}C) from equilibrium response experiments with an atmospheric GCM with a mixed-layer ocean coupled to a tropospheric chemistry model, forced with present-day atmospheric concentrations of CO_2 (a) and by the combined effects of present-day CO_2 levels and sulphur emissions (b). Observed temperature changes from 1955-74 to 1975-94, shown in Figure 9, are qualitatively more similar to the changes in the combined forcing experiment than in the CO_2 only experiment.

Figure 18: Projected global mean surface temperature changes from 1990 to 2100 for the full set of IS92 emission scenarios. A climate sensitivity of 2.5°C is assumed.

Figure 19: Projected global mean surface temperature change extremes from 1990 to 2100. The highest temperature changes assume a climate sensitivity of 4.5°C and the IS92e emission scenario; the lowest a climate sensitivity of 1.5°C and the IS92c emission scenario and the mid-range curve a climate sensitivity of 2.5°C and the IS92a Scenario. The solid curves include the effect of changing aerosol; the dashed curves assume aerosol emissions remain constant at their 1990 levels.

Figure 20: Projected global mean sea level rise from 1990 to 2100 for the full set of IS92 emission scenarios A climate sensitivity of 2.5°C and mid-value ice melt parameters are assumed.

Figure 21: Projected global mean sea level rise extremes from 1990 to 2100. The highest sea level rise curve assumes a climate sensitivity of 4.5°C, high ice melt parameters and the IS92e emission scenario; the lowest a climate sensitivity of 1.5°C, low ice melt parameters and the IS92c emission scenario and the middle curves a climate sensitivity of 2.5°C, mid-value ice melt parameters and the IS92a Scenario.

Figure 22: The pattern of surface temperature change projected at the time of CO_2 doubling from a transient coupled model experiment.

Figure 23: The pattern of surface temperature change projected by a transient coupled model at the time of CO_2 doubling when both CO_2 and aerosol concentration increases are taken into account.

Figure 24: The global mean surface temperature response to the CO_2 concentration pathways leading to stabilisation at 450 (dashed curves) and 650 (solid curves) ppmv (see Figure 7a) for a climate sensitivity of 1.5, 2.5 and 4.5 °C. The changes shown are those arising from CO_2 increases alone. The date of concentration stabilisation is indicated by the dot. Calculations assume the "observed" history of forcing to 1990, including aerosol effects and then CO_2 concentration increases only beyond 1990.

Figure 25: The global mean sea level response to the CO_2 concentration pathways leading to stabilisation at 450 (dashed curves) and 650 (solid curves) ppmv (see Figure 7a) for a climate sensitivity of 1.5, 2.5 and 4.5 °C. The changes shown are those arising from CO_2 increases alone. The date of concentration stabilisation is indicated by the dot. Calculations assume the "observed" history of forcing to 1990, including aerosol effects and then CO_2 concentration increases only beyond 1990.