

GREENHOUSE GASES AND GLOBAL WARMING

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Contents

1. Introduction
 2. Greenhouse Effect
 3. Greenhouse Gases
 4. Radiative Forcing of Climate Change
 5. Non Greenhouse Gas Radiative Forcing
 6. Climate Models
 7. Predictions of Future Climate
 8. Conclusions
- Glossary
Bibliography
Biographical Sketches
To cite this chapter

Summary

Global warming is an important environmental issue which is rapidly becoming a part of popular culture. This paper provides an account of the science associated with this important issue. Historical evidence for past climate change is discussed. The difference between weather and climate is highlighted. The physics of the greenhouse effect and the concept of greenhouse gases are presented. The concepts of radiative forcing of climate change and global warming potential as measures of the absolute and relative strengths of greenhouse gases are discussed. Global warming, the enhancement of the natural greenhouse effect caused by emissions associated with human activities of greenhouse gases such as carbon dioxide, methane, nitrous oxide, and halogenated compounds (e.g. CFCs and SF₆), is described. Techniques used to model past, current, and future climate are discussed. The models are based upon fundamental well established scientific principals and incorporate the current understanding of the complex feedback and couplings between the atmosphere, hydrosphere, and biosphere.

Projections of future global climate change from state-of-the-art computer models are given. With our current level of scientific understanding we expect that over the next century the world will warm substantially.

1. Introduction

Weather and climate have a profound impact on living organisms on the planet. Ecological systems have evolved over geological time scales to suit the prevailing climate. The past 10 to 20 years have brought disturbing evidence that human activities may cause significant changes in future global climate. "Global Warming" is now an issue known to hundreds of millions of people across the world. We provide herein an overview of the current state of knowledge concerning greenhouse gases and global warming. For detailed scientific reviews of the subject with numerous references to the latest technical articles the reader should consult the comprehensive IPCC reports on the subject listed in the Bibliography. For a thought provoking historical perspective we recommend the reader consult the articles on the greenhouse effect published by Fourier in 1827 and Arrhenius in 1896. Our purpose here is to address the following questions: "Is global climate changing?", "To what degree are human activities responsible for climate change?", "What is the state of the science?", "What will future global climate be like?" and "What should we do?".

At this point it is germane to note the difference between weather and climate. Weather is the state of the atmosphere (temperature, humidity, precipitation, wind, cloud cover, etc.) in a particular location at a particular time; it fluctuates greatly and is notoriously difficult to predict. Climate is the time-averaged weather in a given geographical region. Climate is a temporal and spatial average and is consequentially much more predictable than weather. Thus, the average temperature during a given month in a particular area (climate) can be predicted with some confidence, however, the temperature at a given time and location (weather) is much more difficult to predict. Climate varies from month to month, season to season, and year to year. Statistically significant changes in climate occurring over a time scale of decades or longer constitute "climate change".

Climate change is not a new phenomenon in the Earth's history. The geological record shows that when viewed over a time scale of thousands of years the climate is in a state of more or less continual change with major ice ages occurring approximately every 100 000 years. Regular occurrence of ice ages has an astronomical origin associated with subtle changes in the separation and relative orientation of the Earth and the Sun. Three regular variations occur as the Earth orbits around the Sun. First, the Earth's orbit is slightly elliptical and the eccentricity (degree to which it differs from circular) varies with a period of approximately 100 000 years. Second, the Earth's rotational axis is tilted with respect to its orbital axis, the tilt oscillates between 21.6° and 24.5° with a period of approximately 41 000 years (currently the tilt is 23.5°). Third, the month when the Earth is at perihelion (closest approach) with respect to the Sun changes with a period of approximately 23 000 years. The three regular variations give rise to latitudinal and seasonal variations of solar radiation intensity (up to 10% variation in polar regions in summer time) which provides an astronomical forcing of climate leading to ice ages in which the global average temperature is believed to be approximately 5-10 $^{\circ}\text{C}$ cooler than at present. Currently, we are in a period between two

ice ages, in the absence of other influences we will enter another ice age over a time scale of thousands of years. The rate of global temperature change associated with astronomical forcing is believed to be of the order of a few tenths of a degree Celsius per thousand years.

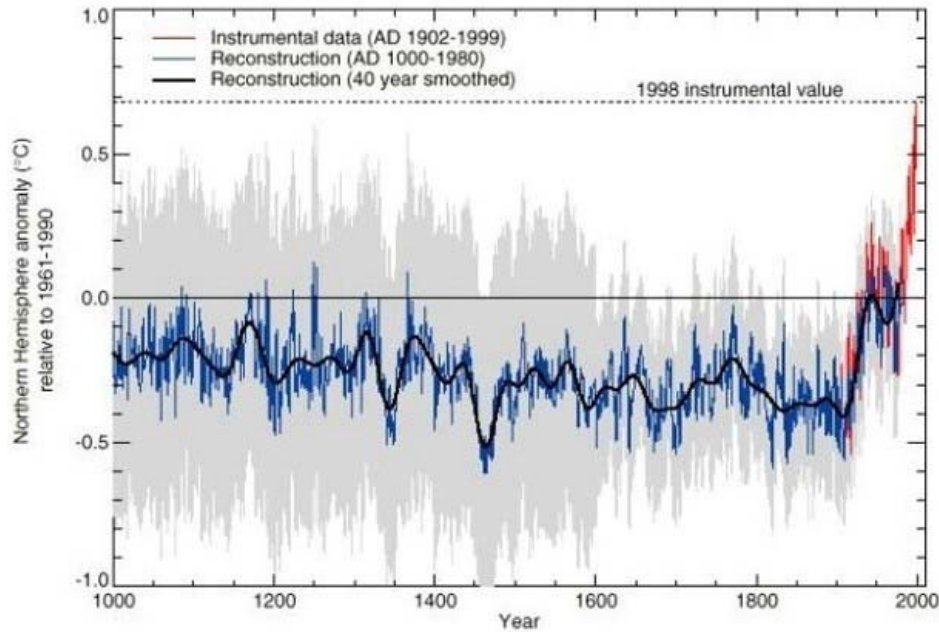


Figure 1. Average Northern Hemisphere temperature over the past 1000 years. Red = instrumental record, blue = reconstructed using proxy indicators, black = 40 year average, grey = estimated uncertainty range. (Reproduced with permission from IPCC)

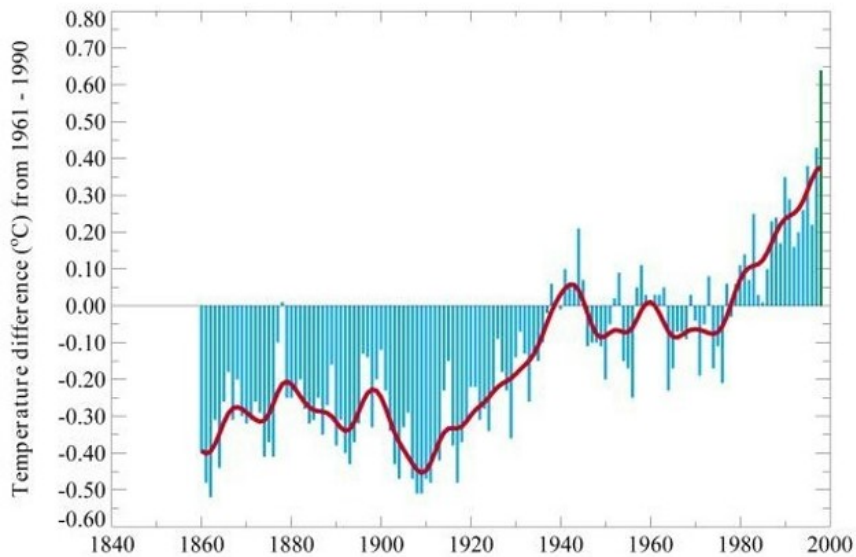


Figure 2. Global average temperature since 1860 (reproduced with permission from IPCC)

Figures 1 and 2 show the average annual Northern Hemisphere surface temperature over the past 1000 years and the average global temperature over the past 140 years. The red

line in Figure 1 is the instrumental record which stretches back approximately 140 years, the blue line is the "reconstructed" temperature deduced from tree rings, ice cores, corals, and historical records. The black curve is the 40 year smoothed average. The temperature record for the Southern Hemisphere is less complete, but is consistent with, that of the Northern Hemisphere. Several interesting points arise from inspection of Figure 1. First, the average Northern hemispheric temperature is far from constant over the past 1000 years. Many fluctuations are observed. Second, there is a small but discernable general trend of decreasing temperature over the time period 1000–1900 with a magnitude consistent with that expected from astronomical forcing. Third, and most significant for the present discussion, there is a pronounced and abrupt (in geological terms) warming over the past century. Figure 2 provides a more detailed picture over the past 140 years which shows that the warming occurred during two periods (1910–1940 and 1980–present). The vertical bars in Figures 1 and 2 represent the difference between the average annual temperature for each year and the average for 1961–1990. 1998 was the warmest year, the 1990s was the warmest decade, and the 1900s was the warmest century in the past millennium. The data in Figures 1 and 2 indicate that global climate is changing.

The temperature record is not the only indication of a changing climate. There are many other indicators such as the substantial retreat of mountain glaciers in many locations around the world, decreased snow cover in the Northern hemisphere, decreased tropical precipitation, increased mid-to-high latitude precipitation, sea level rise, decreased extent of Arctic ice, and thinning of Arctic ice. The combined data prove beyond a reasonable doubt that global climate is changing.

Having concluded that global warming is indeed occurring, the next question is "Should we be concerned?" The warming observed thus far (0.6 ± 0.2 °C since the late nineteenth century) is modest and, in itself is unlikely to lead to a substantial global impact. Concern regarding global warming centers on the future climate. As discussed in detail below, it seems likely that most of the warming observed to date is associated with human activities releasing greenhouse gases such as carbon dioxide. In light of the long atmospheric lifetime of carbon dioxide, substantial increases in its emission during the past century, and expected increases in future emissions, the atmospheric concentration of carbon dioxide in the year 2100 will almost certainly be substantially greater than that today and the climate will continue to warm. The IPCC has considered numerous emission scenarios and concluded that global temperature may rise between 1.4 and 5.8 °C by 2100 (depending on emission scenario). It is instructive to consider Figure 1 and extrapolate a 1.4–5.8 °C increase by 2100, it is clear that when compared to natural variability over the past 1000 years an additional 1.4–5.8 °C change in the next 100 years would be very significant.

2. Greenhouse Effect

The history of the scientific understanding of the greenhouse effect can be traced back to Jean Baptiste-Joseph de Fourier in 1827. Fourier suggested that the atmosphere allows short wavelength radiation from the Sun to reach and warm the Earth's surface but blocks emission of longer wavelength radiation which would cool the Earth. Fourier compared this effect to that of the glass in a greenhouse (hence "greenhouse effect") and

predicted that climate could be influenced by human activities. It is now known that prevention of the loss of warm air via convection is the main factor by which heat is retained in a greenhouse. However, for historical reasons, the terms "greenhouse effect" and "greenhouse gases" are still used when describing radiative trapping of heat in the atmosphere. In 1859 John Tyndall measured the infra-red (heat) trapping ability of water vapor, carbon dioxide, and methane and suggested that ice ages were caused by variations of the atmospheric levels of these gases. In 1896 Svante Arrhenius noted the release of large amounts of carbon dioxide by the burning of fossil fuel and estimated that a doubling of atmospheric carbon dioxide would warm the Earth by 5-6 °C (current climate models predict a 1.5 – 4.5 °C rise from doubling of carbon dioxide). In 1957 Roger Revelle and Hans Suess pointed out that the build-up of carbon dioxide in the atmosphere constituted a large scale geophysical experiment whose consequences were unknown. The following year (1958) was designated as the International Geophysical Year and saw the start of an ongoing program of continuous measurements of atmospheric carbon dioxide levels at Mauna Loa, Hawaii, by Charles Keeling which have showed that the levels of carbon dioxide are rising steadily (from 315 ppm in 1958 to 370 ppm in 2001).

It has been known for over a hundred years that water vapor, carbon dioxide, and methane naturally present in the atmosphere trap heat in the atmosphere. It is relatively straightforward to estimate the magnitude of this "natural greenhouse effect". Sunlight arriving at the top of the atmosphere has a power of 1370 W m⁻² (Solar Constant). The Earth presents a disc of area πr^2 (r = radius of Earth) towards the Sun and as the Earth rotates the sunlight is distributed over a surface area of $4\pi r^2$. Hence, the 24 hour average sunlight at the top of the atmosphere has a power of $1370/4 = 342$ W m⁻². Scattering by air molecules and reflection from surfaces (e.g. by clouds and snow) return an average of 107 W m⁻² back to space leaving 235 W m⁻² to heat the Earth. The only mechanism by which the Earth can cool itself is via emission of infrared radiation into space. At infrared frequencies the Earth behaves like a black body (when seen in IR light even "white" clouds and snow appear "black"). Emission from a black body is given by $\sigma \times T^4$, where σ is the Stefan-Boltzmann constant (5.67×10^{-8} J m⁻² K⁻⁴ s⁻¹) and T is the temperature of the black body. Hence, at radiative equilibrium we would expect the average surface temperature to be approximately 254 K (-19 °C). In reality the average surface temperature of the Earth is 288 K (15 °C); the 34 °C difference is attributable to trapping of infrared radiation (heat) in the atmosphere by greenhouse gases, i.e. the natural greenhouse effect. Without the natural greenhouse effect the planet would be permanently frozen and devoid of life. "Global warming" refers to the enhanced greenhouse effect expected to result from an increase in atmospheric concentration of greenhouse gases resulting from emissions associated with human activities.

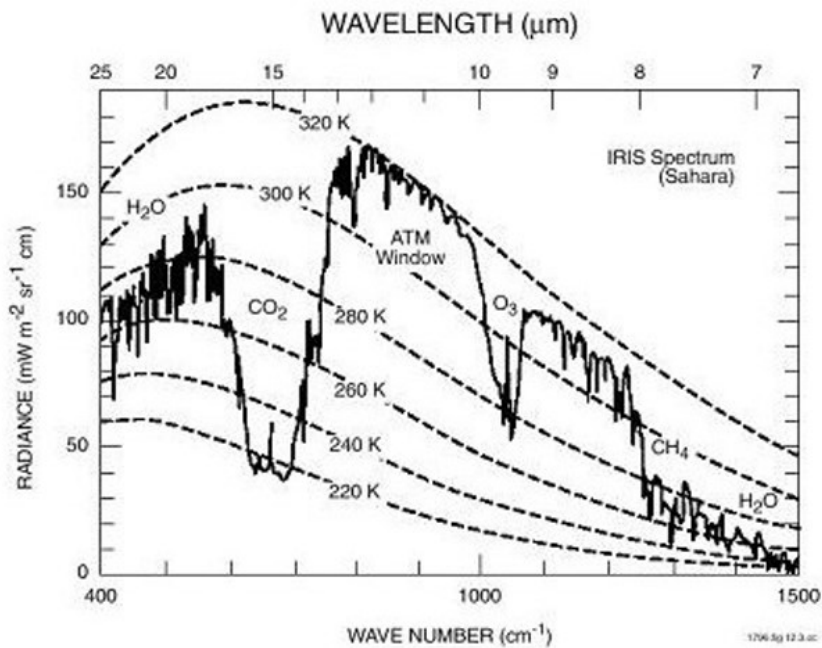


Figure 3. Terrestrial infrared radiation spectrum recorded by an orbiting satellite over the Sahara desert. (Reproduced with permission from Oxford University Press)

Absorption of infrared irradiation by greenhouse gases in the atmosphere has been well documented by satellite observations. Figure 3 shows the infrared radiation observed by a satellite looking down on the Sahara Desert. The dotted lines show emission spectra expected from blackbodies at various temperatures. The atmosphere is transparent at wavelengths of 10-12 μm ($800\text{-}1000\text{ cm}^{-1}$), infrared radiation at this wavelength reaching the satellite comes directly from the surface of sand grains in the Sahara which have a temperature of 320 K ($47\text{ }^\circ\text{C}$). The 10-12 μm region is known as the atmospheric window region. It can be seen from Figure 3 that the infrared radiation on either side of the atmospheric window is characteristic of emission from a blackbody which is significantly cooler than 320 K. Thus, for example the intensity of radiation at wavelengths around 15 μm is characteristic of that expected from a blackbody at approximately 220 K. The apparent discrepancy between the temperature inferred from the 10-12 μm and 15 μm radiation is explained by the fact that carbon dioxide absorbs strongly at 15 μm . Hence, at 15 μm the satellite does not observe emission from the Earth's surface but instead observes emission from high in the atmosphere where the temperatures are much colder than at the surface and the emission much weaker. Water vapor, ozone, and methane are also strong absorbers of infrared radiation. As indicated in Figure 3, the presence of these greenhouse gases in the atmosphere prevents the direct escape of terrestrial infrared radiation into space. Increases in the atmospheric concentrations of greenhouse gases such as carbon dioxide, water vapor, methane, and ozone will act to further hinder the escape of terrestrial infrared radiation and hence warm the Earth's surface. Certain industrial gases such as sulfur hexafluoride (SF_6) and the chlorofluorocarbons (CFCs) are powerful absorbers in the atmospheric window region and have a particularly pronounced greenhouse effect (on a mass basis SF_6 is 22 200 times more powerful a greenhouse gas than carbon dioxide). To understand the

potential for human induced global warming we need to consider the human contribution to greenhouse gas levels.

3. Greenhouse Gases

The three most powerful long lived greenhouse gases in the atmosphere are carbon dioxide, methane, and nitrous oxide. In this section the sources, sinks, and atmospheric concentrations of these compounds are considered. In addition we will consider the class of compounds known as halogenated organic compounds (of which CFCs are a subset), SF₆, and ozone in the lower and upper atmosphere. At this point it is germane to note that water vapor is the single most powerful greenhouse gas in the atmosphere. Water vapor has approximately twice the effect of the second most powerful greenhouse gas—carbon dioxide. Human activities do not have any significant direct impact on the level of water vapor in the atmosphere. However, as a result of global warming it is likely that human activities will have a significant indirect impact on the level of water vapor in the atmosphere. Water vapor is the most important greenhouse gas and the development of a better understanding of the effect of global warming on atmospheric water in all its forms (solid, liquid, and gas) is of critical importance.

Figure 4 shows the concentrations of carbon dioxide, methane, and nitrous oxide in the atmosphere over the past 1000 years as measured in air trapped in bubbles in ice core samples from sites in Antarctica and Greenland and, for the more recent data, measured directly in ambient air. The units for carbon dioxide concentration are parts per million (ppm) while those for methane and nitrous oxide are parts per billion (ppb), 1 ppm = 1000 ppb. The current atmospheric concentration of carbon dioxide is 370 ppm, meaning that in samples of one million molecules of ambient air we would expect to find, on average, 370 molecules of carbon dioxide. As seen from Figure 4, the overall trend of carbon dioxide, methane, and nitrous oxide concentrations is the same. For the time period 1000–1800 there is little, or no, discernable change, while over the period 1800–2000 there is a substantial increase in the atmospheric levels of all three gases. It is well established that the increases in the atmospheric concentration of carbon dioxide, methane, and nitrous oxide during 1800–2000 reflect the impact of human activities in general and the industrial revolution in particular.

3.1. Carbon Dioxide (CO₂)

Discussion of the human impact on the levels of carbon dioxide (CO₂) in the atmosphere is complicated by two factors. First, emissions of CO₂ associated with human activities, while large on a human scale, are small when compared to natural fluxes of CO₂ associated with photosynthesis, respiration, uptake into ocean water, and release from ocean water. Second, there are several large reservoirs of CO₂ (e.g. atmosphere, upper ocean, deep ocean, biosphere) which are continually exchanging CO₂. In such a system one needs to be very careful when using the words "source", "sink", and "lifetime". Instead of "source" and "sink" it is often better to refer to "net source" and "net sink". Thus, the oceans are both a large (90 GtC yr⁻¹) source and a large (92 GtC yr⁻¹) sink of atmospheric CO₂. Overall the oceans provide a net sink for CO₂ of 2 GtC yr⁻¹. The unit used to measure CO₂ flux is GtC which stands for Giga (10⁹) tonnes of carbon. The molecular weight of CO₂ is 44 while the atomic weight of

carbon is 12, hence the mass of CO_2 corresponding to 1 GtC is actually $(44/12) = 3.7$ Gt. With continual exchange between reservoirs it is not possible to specify a single atmospheric lifetime for CO_2 . The atmospheric lifetime of CO_2 is typically quoted as approximately 100 years. For simplicity the discussion below will not consider the natural fluxes of carbon between the various reservoirs in the environment; instead it will focus on the human perturbation to the natural cycle.

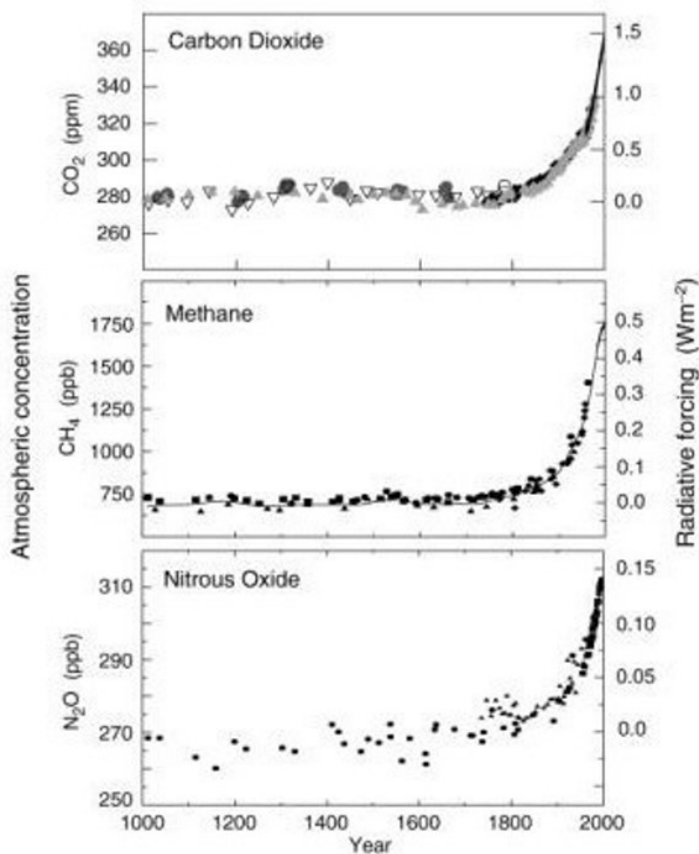


Figure 4. Atmospheric concentration of carbon dioxide (CO_2), methane (CH_4), and nitrous oxide (N_2O) over the past 1000 years. Different symbols are measurements using ice cores from different sites in Antarctica and Greenland. The straight line in the CO_2 plot are the recent in-situ atmospheric measurements. The curve through the CH_4 data is a smoothed fit. (Reproduced with permission from IPCC)

Four approaches have been used to estimate the net CO_2 flux between atmosphere, oceans, and terrestrial biosphere: (i) direct local flux measurements which are extrapolated globally (ii) indirect assessment using atmospheric models calibrated, or partially validated, using various tracers (iii) monitoring of tracers coupled to the CO_2 cycle such as ratios of molecules containing different carbon isotopes, e.g. $^{13}\text{CO}_2/^{12}\text{CO}_2$, $^{14}\text{CO}_2/^{12}\text{CO}_2$, and (iv) measurements of the change in atmospheric CO_2 concentrations as air passes over a particular region combined with "inverse" modeling to estimate the fluxes necessary to account for the observed changes. Such techniques have been used to estimate that the oceans remove $2.0 \pm 0.8 \text{ GtC yr}^{-1}$ and re-growth of forests in the Northern Hemisphere removes $0.5 \pm 0.5 \text{ GtC yr}^{-1}$ of CO_2 from the atmosphere. Human

activities are believed to lead to emission of $5.5 \pm 0.5 \text{ GtC yr}^{-1}$ from fossil fuel combustion and cement production and emission of $1.6 \pm 1.0 \text{ GtC yr}^{-1}$ from changes in tropical land use (deforestation). The atmospheric burden of CO_2 is increasing at a rate of $3.3 \pm 0.2 \text{ GtC yr}^{-1}$. To balance the CO_2 budget "unknown terrestrial sinks" have been invoked and have been inferred to account for $1.3 \pm 1.5 \text{ GtC yr}^{-1}$ of CO_2 . This is often known as the "missing sink". It should be noted that zero lies within the range 1.3 ± 1.5 and hence there may indeed be no "missing sink" (*see Oxides of Carbon*).

It should be noted that a highly simplified picture has been presented above. In reality the carbon cycle is very complex with a number of important feedback loops. One direct carbon cycle feedback is that certain agricultural and wild plant and tree species have shown significantly higher growth under higher (typically doubled) carbon dioxide concentrations. Growth is also strongly influenced by nutrient and water availability as well as other stress factors (temperature, insect attacks etc.). For terrestrial systems CO_2 -climate feedbacks include effects of temperature, precipitation and radiation changes (changes in cloudiness) on primary production and decomposition. For marine systems feedback occurs through climatic influences on ocean circulation and chemistry. Rates of biological activity generally increase with warmer temperatures and increasing moisture. The feedbacks of changing carbon dioxide and climate on ecosystems are many and complex and will not be discussed here.

The current (2001) atmospheric carbon dioxide concentration is 370 ppm and increasing at approximately 1.5 ppm per year. As shown in Figure 4 this level is 30% above that observed in pre-industrial times (280 ppm). The ice core record from Vostok in Antarctica shows that the current level of carbon dioxide in the atmosphere is substantially greater than at anytime in at least the past 420 000 years.

3.3. Methane (CH_4)

Methane (CH_4) is the most abundant well mixed greenhouse gas after carbon dioxide. In contrast to carbon dioxide, methane is removed from the atmosphere via chemical reaction with hydroxyl (OH) radicals. Methane plays an important role in atmospheric chemistry and it can influence the levels of other important trace species via its reaction with OH. All other factors being constant, increased atmospheric levels of methane will result in decreased concentrations of OH and hence a longer lifetime for any gas whose atmospheric lifetime is influenced by reaction with OH. Also, an increase in methane will lead to the production of more tropospheric ozone which is an important greenhouse gas. Methane is involved in complex feedback loops in atmospheric chemistry (*see Hydrocarbons in the Atmosphere and Gas-Phase (Photo-)Chemical Processes in the Troposphere*).

Methane is emitted into the atmosphere by a large number of natural and anthropogenic sources. Natural sources are believed to contribute approximately 30% of the methane flux while anthropogenic sources account for the remaining 70%. Natural sources are estimated to contribute a total of approximately $160 \text{ Tg}(\text{CH}_4) \text{ yr}^{-1}$ ($1 \text{ Tg} = 10^{12} \text{ g}$, $1000 \text{ Tg} = 1 \text{ Gt}$). The largest natural sources are wetlands, termites, and oceans which emit 115, 20, and $10 \text{ Tg}(\text{CH}_4) \text{ yr}^{-1}$, respectively. Anthropogenic sources are natural gas facilities, coal mines, petroleum industry, coal combustion, enteric fermentation, rice paddies,

biomass burning, landfills, animal waste, domestic sewage and are estimated to emit 40, 30, 15, 15, 85, 60, 40, 40, 25 and 25 Tg(CH₄) yr⁻¹, respectively, for a total anthropogenic contribution of 375 Tg(CH₄) yr⁻¹. The identified sources total approximately 535 Tg(CH₄) yr⁻¹.

Methane is removed from the atmosphere through reaction with OH radicals in the troposphere and stratosphere with rates estimated to be 445 and 40 Tg(CH₄) yr⁻¹, respectively. The tropospheric lifetime with respect to the OH radical reaction is approximately 12 years. CH₄ fluxes in the terrestrial and marine systems are highly variable; the soil sink is estimated to 30 Tg(CH₄) yr⁻¹. The sinks for methane total approximately 515 Tg(CH₄) yr⁻¹. Over the past 25 years the annual growth rate of the atmospheric concentration of methane has varied substantially between 0 and 15 ppb yr⁻¹, corresponding to increases in the atmospheric burden of between 0 and 40 Tg yr⁻¹. There is no quantitative understanding of the variation in rate of increase in the atmospheric methane concentration over the past few decades. As seen from Figure 4, it is clear that during the time period 1000 – 1750 there was little, or no, change in the global atmospheric methane concentration, but since 1750 there has been a substantial (150%) increase in the concentration of this greenhouse gas.

3.4. Nitrous Oxide (N₂O)

Nitrous oxide (N₂O) is the third most abundant well mixed greenhouse gas after carbon dioxide and methane. N₂O is a long-lived (130 years) trace constituent of the lower atmosphere present in a concentration which is currently 313 ppb and increasing at a rate of 0.5-0.9 ppb yr⁻¹. As seen from Figure 4, the atmospheric concentration of N₂O has increased by approximately 16% since pre-industrial times. In addition to its importance as a greenhouse gas, N₂O is transported through the troposphere into the stratosphere where it reacts with O(¹D) atoms and is the source of stratospheric NO_x (O(¹D) atoms are electronically excited oxygen atoms). Natural sources of N₂O associated with emission from soils and the oceans are estimated to deliver 10.2 TgN yr⁻¹ to the atmosphere. Anthropogenic emissions of N₂O are associated with biomass burning, fossil fuel combustion, industrial production of adipic and nitric acids, and the use of nitrogen fertilizer, and are believed to total 3.2 TgN yr⁻¹. Photodissociation in the stratosphere is the major (90%) loss mechanism for N₂O in the atmosphere. Reaction with O(¹D) atoms is a minor (10%) loss of N₂O (*see Reduced and Oxidized Nitrogen*).

3.5. Halogenated Organic Compounds

Halogenated organic compounds are organic compounds containing one or more halogen atoms. Halogenated organic compounds can be fully substituted where all of the hydrogens in the molecule have been replaced by halogen atoms, or partially substituted where some hydrogens remain. Chlorofluorocarbons (CFCs) and perfluorocarbons (PFCs) are two subsets of halogenated organic compounds in which all hydrogen atoms have been substituted by fluorine and chlorine atoms, or solely by fluorine atoms. CFC-12 (CCl₂F₂) and CFC-11 (CCl₃F) are the two most abundant CFCs in the atmosphere and are present at levels of 0.5 and 0.25 ppb, respectively. CF₄ is the most abundant PFC and is found at a concentration of 0.08 ppb in the atmosphere. Hydrofluorocarbons (HFCs) and hydrofluorochlorocarbons (HCFCs) are compounds in

which some, but not all, of the hydrogens have been replaced with fluorine (HFCs) or fluorine and chlorine atoms (HCFCs). Halons are a class of compounds containing bromine and chlorine (but no hydrogen). There are no significant natural sources of CFCs, PFCs, HFCs, HCFCs, or Halons. These compounds were not present in the pre-industrial atmosphere, and their presence in contemporary air reflects emissions associated with industrial activities (*see Ecological Chemistry, Chemistry of Organic Pollutants, including Agrochemicals, and Multimedia Fate and Transport of Organic Pollutants*).

Carbon-halogen bonds (e.g. C-F, C-Cl, C-Br) absorb strongly at infrared wavelengths and this is why halogenated organic compounds are strong greenhouse gases. The effectiveness of these compounds as greenhouse gases depends on two factors: (i) the number of carbon-halogen bonds in the molecule and (ii) the atmospheric lifetime of the molecule. In general, the atmospheric lifetime of organic compounds is dictated by their reactivity towards OH (hydroxyl) radicals. Hydroxyl radicals are essentially unreactive towards C-F, C-Cl, and C-Br bonds and hence increasing halogenation of the molecule increases the atmospheric lifetime and the greenhouse strength of the molecule. PFCs have lifetimes of the order of thousands of years while CFCs and Halons persist for typically 50-100 years. By contrast HFCs and HCFCs have relatively short atmospheric lifetimes ranging typically from a few months to a decade (depending on the particular compound). Recognition of the adverse environmental impact of CFCs and Halons on stratospheric ozone has led to international controls on the use of these chemicals. Under the Montreal Protocol the production of CFCs and Halons are being phased out and these compounds are being replaced by HFCs and HCFCs. While release of CFCs and Halons from existing equipment continues, the rate of such release has decreased substantially and the atmospheric concentration of such compounds is beginning to decline. With accelerating use of HFCs and HCFCs over the past 10-20 years the levels of these compounds in the atmosphere have increased substantially, HCFC-22 (CHF_2Cl) is the most abundant at a concentration of 0.1 ppb, while HFC-23 (CF_3H), HCFC-141b (CFC_2CH_3), HCFC-142b (CF_2ClCH_3), and HFC-134a (CF_3CFH_2) are each present at a concentration of approximately 0.01 – 0.02 ppb.

3.5. Sulfur Hexafluoride (SF_6)

On a per molecule basis, sulfur hexafluoride (SF_6) is one of the most potent greenhouse gases known. Its potency stems from its intense absorption at $10.3 \mu\text{m}$ (969 cm^{-1}) in the atmospheric window region (see Figure 3) and its extremely long atmospheric lifetime of 3200 years. SF_6 is present in small amounts in fluorites and degassing from these minerals provides a small natural source which results in a natural background concentration of 0.01 ppt. SF_6 is a useful industrial chemical used as an insulating gas in electrical switching equipment. As a result of anthropogenic emissions the current level of SF_6 in the atmosphere is approximately 400 times that of the natural background and increasing at a rate of approximately 0.2 ppt yr^{-1} . Very recently a new SF_6 -like greenhouse gas was detected in the atmosphere: SF_5CF_3 . The level of SF_5CF_3 in the global atmosphere increased from essentially zero in 1960 to 0.12 ppt in 1999. While the concentration of SF_5CF_3 is very low and this compound does not play any significant role in global warming it is of interest because on a per molecule basis it is the most potent greenhouse gas yet identified in the atmosphere. The discovery of

SF₅CF₃ illustrates that there is much that is still to be learnt concerning such greenhouse gases. While it is unlikely that any single major greenhouse gas awaits discovery, the possibility that many compounds such as SF₅CF₃ are present, with each making a small contribution that when summed represent a non-negligible contribution cannot be ruled out at present.

3.6. Ozone (O₃)

Prior to discussing the relationship between ozone and global climate change it is useful to provide a brief background on the atmospheric chemistry of ozone. In contrast to all other greenhouse gases, ozone is not emitted into the atmosphere. Ozone is generated *in-situ* in the atmosphere from two processes: (i) photolysis of molecular oxygen (O₂) which gives oxygen atoms (O) which then add to molecular oxygen to give ozone (O₃) and (ii) oxidation of organic compounds (from natural and man-made sources) in the presence of nitrogen oxides (NO_x). The first process only occurs in the upper atmosphere where there is sufficient short wavelength sunlight to photodissociate molecular oxygen and this process gives rise to the stratospheric ozone layer at altitudes of 20-50 km. The second process occurs throughout the atmosphere but, because of the much greater availability of organic compounds and NO_x near the Earth's surface, is much more important in the lower atmosphere (troposphere). Emission of large amounts of organic compounds and NO_x in urban areas leads to the formation of substantial amounts of ozone in, and downwind of, large metropolitan centers around the world. In discussions of the climatic impact of human perturbations of atmospheric ozone levels a clear distinction must be made between ozone in the upper atmosphere (stratospheric ozone), which has decreased as a result of human activities, and ozone in the lower atmosphere (tropospheric ozone) which has increased as a result of human activities. Finally, in contrast to all other greenhouse gases considered here, the atmospheric lifetime of ozone is short (of the order of days or weeks, depending on local conditions) and hence its concentration responds quickly to changes in atmospheric conditions. Ozone concentrations in the lower atmosphere are typically 10-100 ppb with levels at the low end of the range being characteristic of remote pristine environments and levels at the high end of the range being typical of polluted urban air masses (*see Tropospheric Ozone Pollution, Stratospheric Ozone Depletion*).

As indicated in Figure 5 and discussed in the next section, the depletion of stratospheric ozone during the period 1979-2000 (caused by the release of CFCs and subsequent chemical reactions) has led to a negative radiative forcing of climate which is estimated to have a magnitude of -0.15 W m⁻². International agreements are now in place which should eliminate the emission of CFCs over the next few decades. It is expected that the stratospheric ozone layer will recover during the twenty-first century and that the magnitude of the negative radiative forcing associated with stratospheric ozone loss will decrease with time.

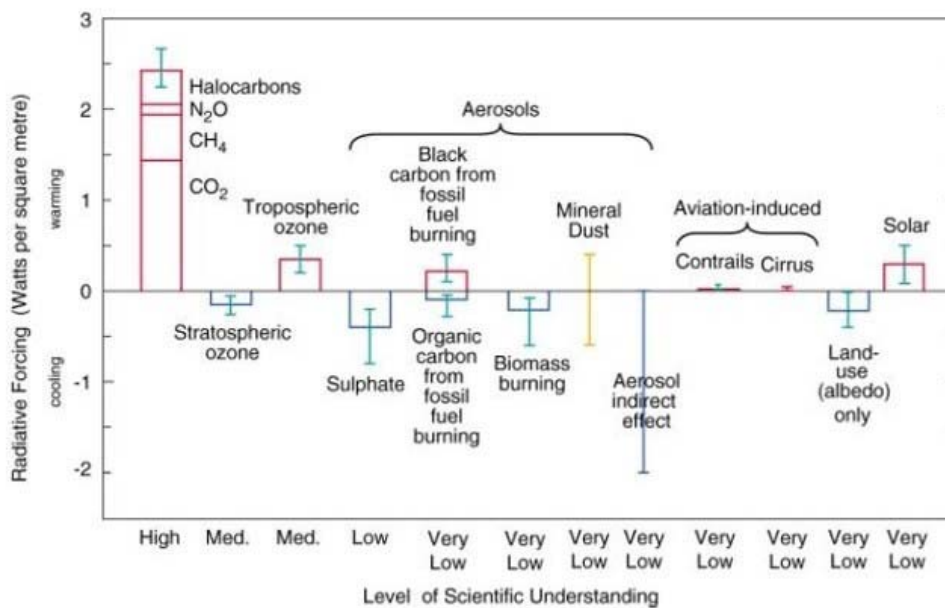


Figure 5. Global mean radiative forcing of climate for the year 2000, relative to 1750. (Reproduced with permission from IPCC)

It is believed that the levels of ozone in the troposphere have increased by 30-40% since 1750 due to increased emission of organic compounds and NO_x. This increased concentration of tropospheric ozone has contributed a positive radiative forcing of 0.35 W m⁻² (see Figure 5). The forcing associated with tropospheric ozone varies substantially by region and season and will respond quickly to changes in emissions of ozone forming compounds.

4. Radiative Forcing of Climate Change

Net radiation is defined as the difference between the solar radiation absorbed by the Earth-atmosphere system and the longwave radiation emitted by the Earth-atmosphere system to space. Net radiation influences the Earth's climate because it determines the energy available for heating the atmosphere, ocean and land. Hence net radiation influences the seasonal variation of rainfall and the strength of the global circulation patterns. When greenhouse gases increase in the atmosphere on account of human activities, the radiative balance of the Earth is altered. The greenhouse gases absorb the longwave radiation emitted by the Earth but are transparent to the radiation coming from the sun. Hence the increase in greenhouse gases causes an increase in the net radiation at the top of the atmosphere. The change in net radiation at the tropopause caused by changes in greenhouse gas or aerosol concentrations is called radiative forcing. Radiative forcing can be calculated accurately if the temperature profile and the concentration of the greenhouse gases in the atmosphere is known. Radiative forcing depends upon how strongly a greenhouse gas absorbs radiation and the location of its absorption bands. As discussed in section 2 and illustrated in Figure 3, the radiation emitted by the Earth-atmosphere system is at its maximum around in the wavelength region 10-12 microns (1 micron = 10⁻⁶ meter). If the absorption band of a greenhouse gas is located in this region then it will tend to have a high radiative forcing. On the other hand, if the absorption lies in a region in which there is already strong absorption

by greenhouse gases presently in the atmosphere (for example at around $15\ \mu\text{m}$ where CO_2 absorbs, see Figure 3) then it will have a low radiative forcing. This later effect explains why increases in atmospheric CO_2 levels have a radiative forcing effect which are small (on a molecule for molecule basis) relative to the effect of other greenhouse gases. During the calculation of radiative forcing one must account for the overlap between the absorption bands of different greenhouse gases. Some gases, such as CFC-12 (CCl_2F_2), can cause direct as well as indirect radiative forcing. The direct radiative forcing of CFC-12 is positive since it has an absorption band in the infrared region. The indirect radiative forcing associated with CFC-12 is negative because emission of CFC-12 leads to the loss of stratospheric ozone with a resulting cooling effect (*see Atmospheric Chemistry and Air Pollution*).

How do we compare the radiative impact of different greenhouse gases released by human beings? Chlorofluorocarbons absorb much more longwave radiation than carbon dioxide on a per molecule basis. The change in atmospheric abundance of chlorofluorocarbons has, however, been much smaller than that of carbon dioxide. To compare the effect of different gases with different abundance and different capability to absorb longwave radiation it is most appropriate to compare their impact on net radiation at the tropopause (boundary between the troposphere and stratosphere, located at approximately 10-15 km). Since pre-industrial times the amount of carbon dioxide in the atmosphere has increased from 278 ppm to 370 ppm and resulted in a radiative forcing change of approximately $1.46\ \text{W m}^{-2}$. The amount of methane in the atmosphere has increased from 700 ppb (in pre-industrial times) to 1700 ppb at present and has resulted in a radiative forcing change of $0.47\ \text{W m}^{-2}$. On the other hand, the amount of CFC-12 has increased from 0 to 0.5 ppb since pre-industrial times and resulted in a radiative forcing change of $0.14\ \text{W m}^{-2}$. Radiative forcing is a measure of the effect of different greenhouse gases or aerosols on radiative balance at the tropopause. Note that the radiative forcing due to increased carbon dioxide is three times as large as that caused by the increase in methane and eleven times as large as that due to CFC-12. Hence, it can be inferred that carbon dioxide is more important than methane or CFC-12 as regards its impact on global warming. It is reasonable to conclude that attempts to significantly reduce global warming should include a reduction of carbon dioxide emissions from human activities. To evaluate the long-term impact of emission of greenhouse gases we need information about the residence times of greenhouse gases emitted on account of human activities. A gas with a long residence time will have a greater impact on future climate than a gas with a short residence time even if the radiative forcing of the two gases is the same. The atmospheric lifetime of carbon dioxide is around 100 years while that of methane is around 12 years. To assist policy makers to understand the potential impact on climate of different greenhouse gases, the concept of Global Warming Potential (GWP) has been introduced. This concept compares the potential impact of different greenhouse gases using carbon dioxide as a reference greenhouse gas. The use of carbon dioxide as a reference greenhouse gas is logical since it is the most important greenhouse gas associated with human activities. Global Warming Potential (GWP) is defined as the ratio of the time integrated radiative forcing from instantaneous release of one kilogram of a substance relative to that of one kilogram of carbon dioxide (CO_2) and can be defined as:

$$GWP(x) = \frac{\int_0^t RF_x [x(t)] dt}{\int_0^t RF_{CO_2} [CO_2(t)] dt} \quad (1)$$

where t is the time horizon (typically 20, 100, or 500 years) over which the integration is performed, RF is the radiative forcing (calculated change in radiation flux at the tropopause) resulting from a unit increase in the atmospheric concentration of a given species and has units of $W m^{-2} ppb^{-1}$, [x(t)] and [CO₂(t)] are concentrations (in units of ppb) of x and CO₂ as a function of time following instantaneous release of 1 kg of each substance. It should be noted that radiative forcing is an absolute measure of the strength of a greenhouse gas on a per volume basis, while GWP is a relative measure on a per mass basis. GWP values for several long-lived greenhouse gases are listed in Table 1. The values given in Table 1 reflect the relative efficiency (on a per mass basis) by which the various gases trap infrared radiation (heat) in the atmosphere. Thus, a given mass of SF₆ over a 100 year time horizon is 22 200 times more efficient at trapping infrared radiation than is the same mass of CO₂. It can be seen from Table 1 that the GWP values change with the time horizon that is considered. The GWPs of some compounds increase with increasing time horizon (e.g. SF₆) while that of others decrease with increasing time horizon (e.g. methane). This different behavior is a reflection of whether the atmospheric lifetimes are longer (e.g. SF₆), or shorter (e.g. methane) than that of CO₂.

Gas	Radiative Forcing ($W m^{-2} ppb^{-1}$)	Atmospheric Lifetime (years)	Global Warming Potential (relative to CO ₂)		
			20 year	100 year	500 year
CO ₂	0.000018	100	1	1	1
CH ₄	0.00037	12	62	23	7
N ₂ O	0.0032	114	284	306	161
CFC-11 (CCl ₃ F)	0.25	45	6,300	4,600	1,600
CFC-12 (CCl ₂ F ₂)	0.32	100	10,200	10,600	5,200
HCFC-22 (CHClF ₂)	0.22	12	5,200	1,900	590
HFC-23 (CF ₃ H)	0.16	257	9,400	12,000	9,900
HFC-134a (CF ₃ CFH ₂)	0.17	14	3,700	1,500	450
SF ₆	0.52	3,200	15,100	22,200	32,400
CF ₄	0.08	50,000	3,900	5,700	8,900
C ₂ F ₆	0.26	10,000	8,000	11,900	18,000

Table 1. Direct Global Warming Potential (GWP) values for selected well mixed greenhouse gases for three different time horizons

Finally, it should be noted that the values in Table 1 are direct GWPs and do not include any indirect effects caused by secondary processes. Some compounds have significant indirect GWPs. For example CFCs deplete stratospheric ozone and hence have a significant indirect impact. The indirect GWP (due to ozone depletion) caused by CFC-

12 over a 20 year time horizon is estimated to be in the range -3100 to -600. Hence, the net GWP due to CFC-12 is estimated to be in the range 7100 to 9600.

5. Non-Greenhouse Gas Radiative Forcing

As discussed in the previous section, changes in greenhouse gases over the past 250 years have produced a positive radiative forcing (2.5 W m^{-2} since 1750) on climate. Research has been focused on these gases as the main culprits of global warming. However, in recent years it has become obvious that many other mechanisms, both natural and anthropogenic, can also produce significant radiative forcings. Figure 5 shows the global mean radiative forcing due to changes in greenhouse gases and these other mechanisms since 1750. As shown in Figure 5, the level of scientific understanding of many of these forcings is well below that of greenhouse gases. We now consider four sources of radiative forcing other than greenhouse gases: aerosols, land use changes, solar variability and volcanic eruptions. Those due to changes in aerosols and land use have anthropogenic components, while the last two are entirely natural forcings.

5.1. Aerosols

Aerosols are tiny sub-micron particles (e.g. soot, mineral dust) or tiny droplets (e.g. sulfuric acid or organic compounds) that are either emitted directly into the atmosphere (primary aerosols) or formed from emitted gases during chemical reactions (secondary aerosols). They result from human activities such as fossil fuel burning, transport emissions, biomass burning and land use changes, but are also produced naturally by vegetation, marine algae, volcanoes and high wind stress over naturally arid regions (*see Atmospheric Chemistry and Air Pollution, Suspended Material / Aerosols, Formation and Effects of Smog*).

Aerosols can affect climate in two main ways. Firstly, they can affect the radiation balance of the Earth—the so-called “direct” effect. Most aerosols reflect incoming solar radiation back to space thereby preventing it reaching the Earth's surface. In this way, increases in the amount of sulfates in the atmosphere have produced a negative forcing of -0.5 W m^{-2} during the past 250 years (as shown in Figure 5). An additional negative forcing has also come from extra droplets of organic compounds from fossil fuel and biomass burning. Some types of aerosol, such as black carbon (soot), also absorb solar radiation, trapping energy in the troposphere. Increases in soot resulting from fossil fuel burning have produced a positive radiative forcing (approximately 0.3 W m^{-2}) since 1750. Increased biomass burning in the past 250 years has caused a radiative forcing of around -0.25 W m^{-2} (biomass aerosol is a combination of sulfates, organic carbon and soot). Larger particles like dust can also absorb infrared radiation emitted by the Earth, preventing it reaching space and therefore acting something like a greenhouse gas. This effect is highly dependent on the surface over which the aerosol lies and so the size and even the sign of radiative forcing due to recent dust changes is uncertain.

Some types of aerosol can also have an “indirect” effect on climate. The particles can act as cloud condensation nuclei, altering the number of droplets in a cloud and therefore its lifetime and brightness. These cloud changes then affect the amount of

radiation reaching the Earth's surface. During the past 250 years, these effects are thought to have resulted in a negative radiative forcing, the magnitude of which could be (on a global mean scale) nearly as large as the total positive greenhouse gas radiative forcing over the same period of time. However, since we know relatively little about the interaction between aerosols and clouds, the final result is very uncertain. Additionally, some aerosols can provide surfaces on which chemical reactions can occur.

In Figure 5 the global mean radiative forcing due to aerosols (except for the indirect effect) appears much smaller than that due to greenhouse gases. This is somewhat misleading because the natures of the two forcings are completely different. Most aerosols can only survive in the atmosphere for a few weeks and so concentrations are only high close to the sources. Thus the radiative forcing is also highly geographically variable (the distribution of radiative forcing due to past changes in sulfate aerosol is shown in Figure 6a). As seen from Figure 6a, sulfate aerosol exerts a large local negative forcing (up to -5 W m^{-2}) over parts of Asia, but exerts essentially no effect over the majority of the Earth's surface. In contrast, carbon dioxide has a long atmospheric lifetime, is mixed more uniformly around the globe, and as a consequence the forcing due to increased carbon dioxide is much more uniform (see Figure 6b).

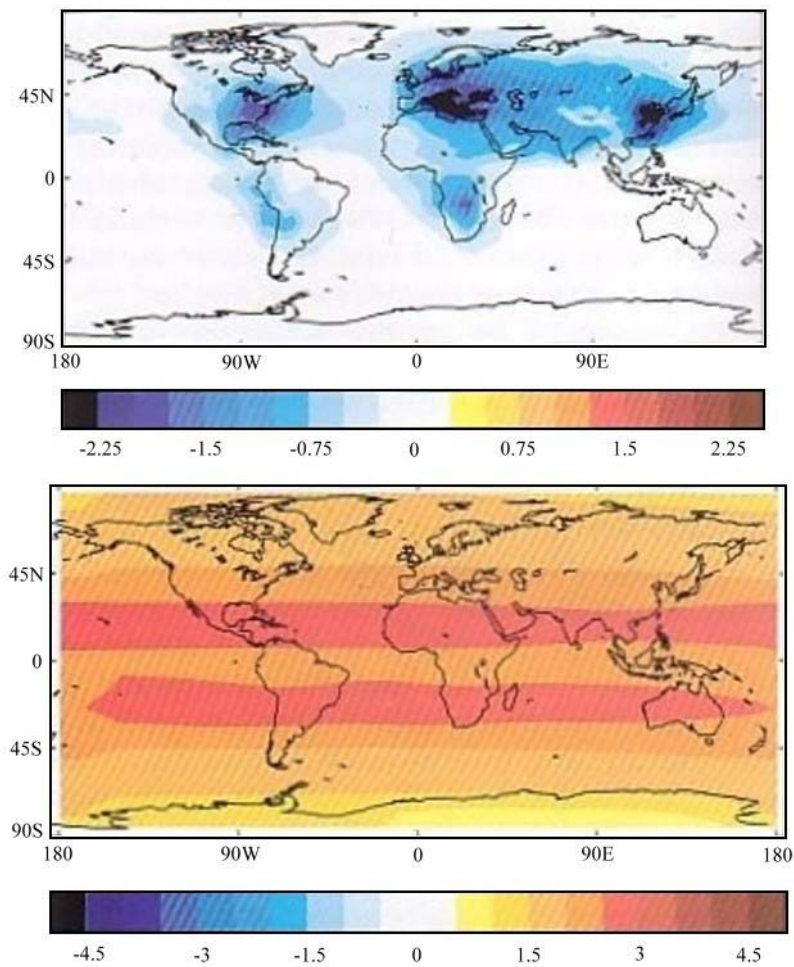


Figure 6. Geographic distribution of radiative forcing due to changes in sulfate aerosol (panel a) and CO₂ (panel b). Adapted from Shine, K.P. and P.M.D. Forster, 1999: "The effect of

human activity on radiative forcing of climate change; a review of recent developments", *Global. Planet. Change.*, 20, 205-225, courtesy of Elsevier Press

The implication here is that it is far too simplistic to think that the effect of aerosols cancels out the positive radiative forcing produced by greenhouse gases. In an analogous fashion a man standing with one foot in boiling water and the other in ice water cannot be said to be on average comfortable. The radiative forcing due to aerosol is highly dependent on type, mixture, relative humidity, size and underlying surface, all of which are currently less than adequately characterized.

5.2. Land Use Changes

Anthropogenic land use changes due to agricultural practices, deforestation, overgrazing, desertification, and urbanization lead to a change in the surface characteristics of the Earth. Over the past 250 years, such changes have led to an increase in albedo (the fraction of incoming radiation returned to space via scattering by gas molecules and reflection by clouds and at the Earth's surface). Over the past 250 years land use changes have produced a negative global mean radiative forcing of -0.2 W m^{-2} (see Figure 5). Human induced land use change may also indirectly cause other radiative forcings, such as an increased atmospheric aerosol loading. These effects currently remain unquantified.

5.3. Solar Variability

Climate can be affected both by changes in the Sun-Earth orbital parameters such as axis tilt, Sun-Earth distance and the obliquity of the orbit. However, these typically vary on timescales of thousands of years and so are generally not thought to have played a major role in climate change over the past 250 years. The total amount of energy emitted by the Sun (or the total solar irradiance, TSI) varies on much shorter timescales, particularly having a prominent 11 year cycle. Periods of higher than usual emission are characterized by the appearance of dark sunspots surrounded by bright faculae. Historically, the Little Ice Age of the fourteenth and fifteenth centuries coincided with the total disappearance of sunspots, indicating that the TSI was very low. It is generally thought that changes in TSI might have been largely responsible for the warming observed in the first half of the twentieth century but could not alone be responsible for the warming in the second half of the twentieth century. Over the past 250 years, solar variability has produced a positive global mean radiative forcing of around 0.3 W m^{-2} .

5.4. Volcanic Activity

Large explosive volcanic eruptions such as that of Mount Pinatubo in the Phillipines during 1991 are capable of putting vast amounts of sulfur dioxide gas into the stratosphere. Here it is converted into sulfuric acid aerosol. Since it is above the weather, this aerosol can remain in the stratosphere for two to three years, and for a tropical eruption, can be spread quickly around the globe by the winds (*see Sulfur Dioxide and Sulfur Cycles*). The aerosol scatters solar radiation back to space (causing the surface to cool) and absorbs some solar and long wave radiation (causing the stratosphere to warm). The Pinatubo eruption produced a peak global mean radiative forcing of around -4 W m^{-2} , and the surface and stratospheric temperatures remained

altered by a few tenths of a degree for up to 4 years after the eruption. Individual eruptions give a transient effect on climate and have limited long term effects. It is therefore difficult to compare their radiative forcing with the other mechanisms shown in Figure 5. However, the cumulative effect of several eruptions during active volcanic periods could produce a prolonged effect on climate. The difference in radiative forcing on comparing an active and quiet decade could be as much as 1 W m^{-2} .

6. Climate Models

How do we predict the impact of increasing greenhouse gases and aerosols on future climate? This can only be achieved using climate models. In atmospheric climate models the entire atmosphere of the Earth is divided into a large number of small boxes. The typical dimensions of these cubes are 200 kilometers in longitude, 200 kilometers in latitude and 1 kilometer in height. Hence a typical climate model has about 400 000 such boxes! The temperature and winds in each box is calculated using the physical laws governing the conservation of mass, momentum, and energy. The moisture content is calculated by ensuring the conservation of water in all its forms (liquid, solid and vapor) in each box. It is not possible to provide analytical solutions to the equations in the models and so a numerical computational approach is used. Since the number of computations involved is very large these climate models require super-computers. The models used for short-term weather forecasting are similar to those used to predict future climate change. For weather forecasting accurate initial values are needed to predict the weather for the next few days. For predicting future climate change, the atmospheric models need information about how the surface temperature of the ocean will evolve. The variation of surface temperature of the ocean can be predicted by ocean climate models. These models are similar to atmospheric climate models and use the laws of conservation of mass, momentum, energy and salinity. If the atmospheric and ocean models are run together they are called coupled models. In such models changes in atmospheric winds will alter the surface temperature of the ocean which in turn will alter the atmospheric winds. These models are run for hundreds of years to predict how the climate of the atmosphere and ocean will evolve together. They have been used to understand how the climate has changed over the last hundred years and predict how they will change during the next hundred years. They are able to accurately simulate the observed increase in global mean temperature during the past hundred years. When these models are used to predict changes in global mean temperature over the next hundred years, they show significant differences. One reason for the variations in model predictions of future climate is the different treatment of clouds. The representation of clouds in climate models is problematic because we are not able to spatially resolve clouds in the models and we lack sufficient observations of the different types of clouds and their physical properties. Hence different modeling groups have represented clouds and their impact on climate in different ways in their climate models.

Over the past decade the sophistication of global scale climate models has increased markedly as our understanding of the complexities of the climate system has grown. This trend is likely to continue. The increasing power of computer systems will make it possible to include more detailed descriptions and interactions between atmosphere, ocean, sea-ice, sulfate aerosol, non-sulfate aerosol, clouds, land use, and atmospheric chemistry in future climate models and will necessitate a much greater understanding of

these processes and phenomena. Moore's Law states that the power of computer systems will double every 18 months. Computer experts believe that Moore's Law will hold over the next 10-20 years. Hence, over the next two decades the computing power available to run global climate models is expected to increase by a factor of 10 000. The observational global climate record received by satellites will improve substantially in both quantity and quality over the coming decades. With the confluence of the availability of dramatically more powerful computer systems, a better climate record to test and calibrate the models, and improved scientific understanding of climate change it seems reasonable to expect that during the next 10-20 years there will be substantial improvements in our ability to predict the likely magnitude, timing, and consequences of global climate change.

7. Predictions of Future Climate

To predict any future event certain assumptions must be made. Predictions of future climate are based upon two main assumptions. First, estimates of future anthropogenic forcing of climate (e.g. emission rates of greenhouse gases, land use changes, aerosol emissions). Second, the validity of climate models used to relate climate to changes in anthropogenic forcings. With regard to both the likely magnitude of future anthropogenic forcing of climate and the validity of climate models it is necessary to rely on expert judgment. The Intergovernmental Panel on Climate Change (IPCC) is widely acknowledged as the most authoritative source of expert judgment in this field (see Bibliography). The IPCC considered greenhouse gas emission scenarios arising from a range of population, economic, cultural, social, and technological trends over the next century and incorporated these scenarios in state-of-the-art climate models. The results from models using 35 different emission scenarios project that the global average annual temperature will increase by 1.4 to 5.8 °C over the period 1990 to 2100 (in addition to the 0.6 °C warming since 1860). Experts at the Massachusetts Institute of Technology have used quantitative uncertainty techniques to assign probabilities to various scenarios and have reported a median value of 2.5 °C for the average global surface temperature rise over the period 1990 to 2100 with a 95% confidence interval of 0.9 - 4.8 °C. In a similar analysis experts at the National Center for Atmospheric Research in Boulder, Colorado, have estimated with 90% confidence that the warming will lie in the range 1.7 – 4.9 °C with a median value of 3.07 – 3.18°C (depending on probability distribution assumptions). With our current level of scientific understanding we expect that in the absence of climate change mitigation policies the global average surface temperature will most probably increase by 2.5 – 3.2 °C over the period 1990 to 2100. To place a global warming of 2.5 – 3.2 °C into perspective, it is instructive to return to Figure 1 and compare this temperature change to that of the natural variability in the climate system. It should be noted that the *entire* y-axis scale in Figure 1 spans 2 °C. A warming of 2.5 – 3.2 °C by 2100 would be very significant. Global climate models project higher maximum temperatures, more hot days, higher minimum temperatures, fewer cold days, fewer frost days, and reduced diurnal temperature variation over essentially all land areas.

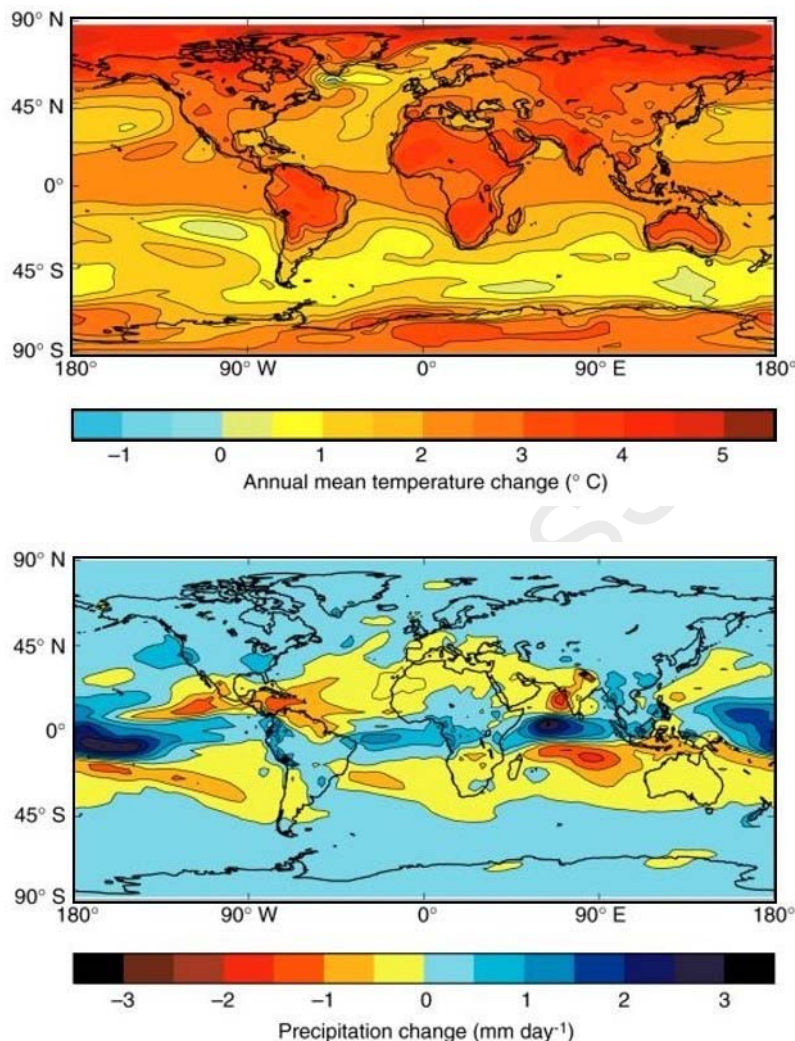


Figure 7. Projected change in average annual temperature (A) and precipitation (B) for the 2050s compared to the present day (assuming an increase in greenhouse gas concentrations equivalent to about a 1% increase per year in CO₂). Reproduced with permission from the Hadley Centre for Climate Prediction and Research, U.K. Meteorological Office

Figure 7 shows the changes in average annual temperature (A) and precipitation (B) over the next 50 years predicted using a state-of-the-art climate model (Hadley Center, U.K. Meteorological Office, see bibliography) assuming an increase in greenhouse gas concentrations equivalent to about a 1% increase per year of carbon dioxide. As seen from Figure 7A, the predicted warming is not geographically uniform. Land areas will warm substantially (up to a factor of 2) more than the global average while ocean areas (with greater thermal inertia) will warm significantly less than the global average. High latitudes (particularly in the Northern Hemisphere) will experience the most intense warming. This effect reflects the importance of the ice-albedo feedback, where the melting of snow and ice will decrease the local albedo (reflectivity) and thereby increase the heating from sunlight. Interestingly, some geographical areas (see Figure 7A) are expected to actually *cool* as a result of global warming. The explanation for this paradoxical result lies in the complexities of the climate system. The North Atlantic receives a substantial amount of heat transported from lower latitudes by the Gulf Stream. Ocean models predict that Global Warming will weaken the Gulf Stream,

reduce the associated heat flow, and cause a cooling trend in certain areas of the North Atlantic.

In a warmer world there will be more evaporation and hence more precipitation. Figure 7B shows changes in precipitation predicted over the next 50 years. While most areas are expected to receive more precipitation than at present, there are large areas which are expected to receive less precipitation. Interestingly, in general, the areas that are predicted to experience reduced precipitation are those in which supplies of fresh water are limited (e.g. Northern Africa, Indian sub-continent, Australia). The global hydrological cycle is expected to be more vigorous with more intense precipitation events. Increased continental drying and increased risk of drought is predicted in mid latitudes.

Prediction of the future global climate is perhaps one of the greatest scientific challenges of the twenty-first century. It is difficult to imagine a more complex or important system than that of climate. Human activities are increasing the levels of important greenhouse gases such as carbon dioxide, methane, and nitrous oxide in the atmosphere to levels which have never been seen before in human history, we are quite literally entering uncharted territory. Climate models are the only available tools to predict future climate. The models are based upon fundamental well established scientific principals and incorporate our current understanding of the complex feedback and couplings between the atmosphere, hydrosphere, and biosphere. With our current level of scientific understanding we expect that over the next century the world will warm substantially.

8. Conclusions

At the beginning of this chapter the following questions were posed: "Is global climate changing?", "To what degree are human activities responsible for climate change?", "What is the state of the science?", "What will future global climate be like?" and "What should we do?". In the light of the discussion in the previous sections we are now in a position to provide answers to these questions.

"Is global climate changing?" The answer to this question is "Yes". There is a large body of scientific data that shows beyond any reasonable doubt that the global climate has warmed significantly over the past 150 years. Probably the most compelling evidence for global warming comes from the temperature record shown in Figures 1 and 2. However, this data is by no means the only indication of a change in climate. In addition, observed retreat of glaciers around the world, decreased snow cover in the Northern hemisphere, decreased tropical precipitation, increased mid-to-high latitude precipitation, sea level rise, increased ocean temperature, decreased extent of Arctic ice, and decreased thickness of Arctic ice all point to a warming global climate. The combined data prove beyond any reasonable doubt that there has been a definite warming trend in global climate over the past century. It is important to stress that the warming trend is in "global climate", not "local climate" or "global weather". Thus, neither a long term cooler climate in restricted geographical areas, nor a decrease in average global temperature from one year to the next should be viewed as inconsistent with the overall warming trend.

"To what degree are human activities responsible for climate change?" The answer to this somewhat controversial question is "It is very likely that human activities are responsible for a substantial fraction of the observed climate change". The natural greenhouse effect is a well known phenomenon. It is well established that human activities have led to large increases in the atmospheric concentrations of the three most important long lived greenhouse gases. State-of-the-art computational climate models are only able to reproduce the historical climate record when forcings associated with human activities are included. Models which ignore forcings associated with human activities are not able to reproduce the historical climate record. Natural climate forcing agents such as the sun and volcanoes play an important role in the climate system but do not explain the observed increase in global temperature over the past century.

"What is the state of the science?" It has been known for a very long time that gases such as water vapor, carbon dioxide, and methane trap infrared radiation in the atmosphere and warm the Earth. In 1827 Fourier compared this effect to that of the glass in a greenhouse from which the term "greenhouse effect" is derived. In 1896 Arrhenius calculated that a doubling of atmospheric carbon dioxide would cause a 5-6 °C increase in global temperature. In 2001, state-of-the-art climate models run on the latest super computers predict that a doubling of carbon dioxide would lead to a 1.5 – 4.5 °C warming. While it is remarkable how little change there has been over the past 105 years in the prediction of the global warming caused by a doubling of carbon dioxide, this fact should not be taken as an indication that the prediction of future climate is a mature scientific discipline. Global climate models are in an early stage of development. There is no doubt that in the coming decades the sophistication and level of detail in the models, and hence their predictive capability, will increase dramatically. Current areas of significant uncertainty in the models are: (i) future levels of emission of greenhouse gases (ii) future levels of atmospheric aerosols (iii) magnitude of feedback mechanisms which amplify (or damp) the climate response to increased greenhouse gas levels (iv) treatment of clouds and (v) the possible existence and importance of metastable climate states. It seems likely that significant surprises lie in store in the future scientific exploration of this fascinating field.

"What will future global climate be like?" As with all questions involving a prediction of future events, this is a difficult question to answer. The best answer to this question is "While the details are unclear, it appears likely that the future global climate will be significantly warmer than that at present (during the next century the global climate will most probably warm by an amount which is approximately 10 times that of the long term natural variability of the climate system indicated by the black line in Figure 1)". There are two fundamental problems associated with prediction of future global climate. The first and most serious problem is that the future emission rates of greenhouse gases are unknown. Will the world continue to rely heavily on fossil fuel for the next century, or will non-fossil fuel energy sources such as nuclear and renewable options gain in importance? It is clear that when looking 10 or 20 years into the future that fossil fuel will continue as the dominant energy source. However, it is very difficult to predict the dominant energy source 50 to 100 years into the future. Global population growth and the development path chosen by the developing nations are important factors affecting future greenhouse gas emissions. The second problem is that even if the future emission rate of greenhouse gases was known exactly, the uncertainties inherent in the scientific

understanding of the link between greenhouse gas emissions and global climate change (see previous paragraph) would still introduce substantial uncertainty in the predictions of future climate. The Intergovernmental Panel on Climate Change considered many different possible emission scenarios for greenhouse gases with a range of different assumptions concerning economic growth, population growth, and reliance on fossil fuels. The projected annual carbon dioxide emissions in the year 2100 from these scenarios varied widely from a high of 30 GtC to a low of 4 GtC. The projected increase in average global temperature for the scenarios is between 1.4 and 5.8 °C with a warming of 2.5 – 3.2 °C being most probable. To place this prediction into context it is useful to refer to Figure 1 and to extrapolate an additional 2.5 – 3.2 °C rise from 2000 to 2100. It is clear that by comparison to the natural variability observed over the past 1000 years, the climate change expected over the next 100 years will be substantial.

"What should we do?" This is of course a value question and as such it is not possible to provide a purely scientific answer. However, science can provide some guidance in addressing this question. Over the past century the demand for energy has increased significantly. This demand has been satisfied primarily by fossil fuel. There was a substantial (12 fold) and sustained (during each and every decade) increase in fossil fuel usage over the past century and it is projected that fossil fuel use will continue to increase for at least the next couple of decades. Carbon dioxide has a long atmospheric lifetime and there is a delay of approximately 100 years between changes in carbon dioxide emissions and equilibration of its atmospheric concentration. In the absence of policies to address the accumulation of greenhouse gases in the atmosphere it appears very likely that over the next century the concentration of atmospheric carbon dioxide will reach a level which is substantially (a factor of 2-3) greater than that of the pre-industrial era. It would seem prudent at this point in time to pursue the research and development of alternative global energy strategies which either decrease our long term reliance on fossil fuels, or reduce carbon dioxide emission associated with fossil fuel use (e.g. geologic sequestering of carbon dioxide). It would also seem prudent to devote additional resources to improve the scientific understanding of the climate system and hence reduce the uncertainties associated with projections of future climate change. Climate change is a long term global issue which requires a long term global solution.

Glossary

Aerosols:	Small solid particles or liquid droplets suspended in the air.
Atmospheric lifetime:	Time taken for a pulse of gas emitted into the atmosphere to decay to 37% (1/e) of its original amount.
Biosphere:	Sum of all living biological systems.
Blackbody:	Surface that absorbs all radiant energy falling upon it. In addition to being a perfect absorber, a blackbody is a perfect emitter of radiation, at a given temperature the radiant energy emitted at each wavelength is the maximum possible.
Carbon dioxide (CO₂):	Colorless gas with sharp odor, formed in combustion of carbon containing materials (e.g. fossil fuels, biomass), in respiration of animals, and fermentation. Taken up by plants during photosynthesis. The main anthropogenic greenhouse gas.

Chlorofluorocarbons:	Industrially made compounds containing chlorine, fluorine, and carbon. Commonly known as CFCs, these non-flammable, non-toxic compounds found widespread use as refrigerants, foam blowing agents, aerosol propellants, and solvents prior to the recognition of their damaging effect on stratospheric ozone.
Climate:	Time-averaged weather in a given geographical region.
Climate change:	Statistically significant change in climate occurring over a time scale of decades or longer.
Climate models:	Complex computer programs incorporating mathematical descriptions of climatically relevant processes in the atmosphere, hydrosphere, and biosphere and their interactions
Global warming:	Warming of climate by enhanced greenhouse effect resulting from an increase in atmospheric concentration of greenhouse gases from anthropogenic emissions.
Global warming potential:	Ratio of the time integrated radiative forcing of a given mass of a well mixed GHG relative to that of the same mass of another well mixed GHG (typically CO ₂) over a specified time horizon. Global warming potential is a useful index to compare the radiative effects of different gases.
Greenhouse effect:	Greenhouse gases absorb infrared radiation passing through atmosphere thereby impeding the flow of heat from the Earth's surface into space. In the early 1800s this effect was compared to that of the glass panes which keep greenhouses warmer than their surroundings – hence "greenhouse effect". As a result of the natural greenhouse effect the Earth's surface is approximately 33 °C warmer than would be the case in the absence of greenhouse gases.
Greenhouse gas:	Gas in the atmosphere that absorbs and emits at wavelengths within the spectrum of infrared radiation emitted by the Earth's surface (terrestrial radiation).
Gt:	10 ⁹ tonnes
Halons:	Compounds containing carbon, bromine, and chlorine which are commonly used as fire fighting agents.
Hydrofluorocarbons:	Compounds containing hydrogen, fluorine, and carbon (HFCs), often used as CFC replacements.
Hydrosphere:	The component of the global climate system comprising liquid water; oceans, lakes, seas, rivers, underground water systems, etc.
Infrared radiation (IR):	Portion of the electromagnetic spectrum that extends from the long wavelength (red) end of the visible spectrum to the microwave region. Invisible to the eye, it produces a sensation of warmth on the skin. IR radiation has frequencies below those of red (infra is Latin for below).
IPCC:	Intergovernmental Panel on Climate Change (see http://www.ipcc.ch/).
Nitrous oxide (N₂O):	Also known as dinitrogen monoxide or laughing gas, an anesthetic. Natural sources include emissions from land and

	ocean; anthropogenic sources include biomass burning, fossil fuel combustion, and industrial processes.
Ozone (O₃):	Triatomic allotrope of oxygen present in troposphere and stratosphere.
ppb:	Parts per billion.
ppm:	Parts per million.
Radiative Forcing:	Change in net vertical irradiance (units of W m ⁻²) at the tropopause caused by internal (e.g. change in GHG or aerosol concentrations) or external change (e.g. change in solar flux) in the climate system.
Solar constant:	Flux of solar radiation arriving at Earth (approximately 1370 W m ⁻²).
Solar radiation:	Electromagnetic radiation emitted from the sun.
Terrestrial radiation:	Infrared radiation from the Earth's surface.
Tg:	10 ¹² g (10 ⁹ kg).
Tonne:	1000 kg.
Troposphere:	Lowest region of the atmosphere (up to approximately 10-15 km) characterized by decreasing temperature with increasing altitude.
Tropopause:	Boundary between the troposphere and stratosphere, located at an altitude of approximately 10-15 km.
Weather:	State of the atmosphere (temperature, humidity, precipitation, wind, cloud cover, etc.) in a particular location at a particular time.

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Biographical Sketches

Timothy J. Wallington was born in England in 1958. He received his B.A. in Natural Sciences (1st Class) in 1981 and his Ph.D. in Physical Chemistry in 1983 from Oxford University under the supervision of Richard Wayne. In 1984 he moved to the Statewide Air Pollution Research Center in Riverside, California, to work with James N. Pitts, Jr. and Roger Atkinson. In 1986 he moved to the National Bureau of Standards (now National Institute of Standards and Technology) near Washington D.C. to work with Michael Kurylo. From 1987 he has worked in the Scientific Research Laboratories of Ford Motor Company in Dearborn to establish the environmental impact of vehicle emissions. In 1998 he spent a year as a Humboldt Research Fellow in Wuppertal, Germany with Karl Heinz Becker. He has co-authored over 300 scientific articles and 2 books in the area of atmospheric science.

Jayaraman Srinivasan was born in India in 1947. He received his B.Tech. in Mechanical Engineering in 1969 from Indian Institute of Technology, Madras, India, his M.S. in Mechanics in 1971 from State University of New York, Stony Brook and his Ph.D. in Mechanical Engineering in 1975 from Stanford University. From 1975 to 1982 he was a faculty member in the Department of Mechanical Engineering at Indian Institute of Technology, Kanpur, India. In 1982 he moved to Indian Institute of Science, Bangalore, India. He is now the Chairman of the Centre for Atmospheric and Oceanic Sciences at Indian Institute of Science. He has done research in Atmospheric Sciences, Thermal Sciences and Solar Energy Utilization. He was a National Research Council Senior Resident Research Associate at NASA, Langley, Virginia from 1993 to 1995. He was one of the lead authors in the Intergovernmental Panel on Climate Change (IPCC) report on Climate Change in 1995. He was review editor of the IPCC report on Climate Change in 2001.

Ole John Nielsen, was born in Denmark in 1954. He received his M.Sc. in Chemistry and Physics in 1978 and his Ph.D. in Atmospheric Chemistry in 1984 from Copenhagen University under the supervision of Børge Bak. From 1978 to 1999 he worked at Risoe National Laboratory (Denmark), 5 years as head of the Section for Chemical Reactivity. In 1995 and 1996 he moved to Ford Research Center Aachen (Germany) where he worked as a manager of atmospheric science. In 1996 he received the Henry Ford Technology Award. In 1997 he was appointed adjunct professor in chemistry at Roskilde University. He is a member of the Danish Natural Science Research Council, the Head of Copenhagen Global Change Initiative (COGCI) network and Ph.D.-board, member of the Danish Academy (ATV), and the Danish member of the International Executive Committee for EUROTRAC. He has supervised numerous M.Sc. and Ph.D. students and has co-authored 145 scientific articles in the area of atmospheric science.

Ellie J. Highwood was born in France in 1972. She received her B.Sc. (Hons) in Physics (1st class) from the University of Manchester in 1993 and her Ph.D. in Meteorology in 1997 from the University of Reading under the supervision of Prof. Brian Hoskins. She then spent 18 months as a research assistant in the same department working on the global warming impact of replacement gases for the chlorofluorocarbons and non-methane hydrocarbons. In 1998 she won a personal Postdoctoral Research Fellowship from the U.K. Natural Environment Research Council to study the climate response to changes in tropospheric aerosols. Since January 2001 she has been a lecturer in the Department of Meteorology at the University of Reading. Her research interests include the radiative and climatic impact of volcanic eruptions, Saharan dust storms and biomass burning. She has co-authored around 15 scientific articles in peer-reviewed publications.

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