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The greenhouse effect and climate change

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Abstract.

On any planet with an atmosphere, the surface is warmed not only by the Sun directly but also by downward-propagating infrared radiation emitted by the atmosphere. On the Earth, this phenomenon, known as the greenhouse effect, keeps the mean surface temperature some 33K warmer than it would otherwise be and is therefore essential to life.

The radiative processes which are responsible for the greenhouse effect involve mainly minor atmospheric constituents, the amounts of which can change either naturally or as a by-product of human activities. The growth of the latter is definitely *tending* to force a general global surface warming, although because of problems in modelling complicated feedback processes, for example those involving water vapour, ozone, clouds, and the oceans, the precise rates of change and the local patterns which should be expected are not yet very well known.

This article reviews the physical processes involved in the greenhouse effect and discusses current progress, theoretical and experimental, towards an understanding of the effect on the climate, especially the mean surface temperature, of recent and expected changes in atmospheric composition. It also provides an overview of recent expert forecasts of climatic change in the next few decades, and discusses their limitations.

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Contents

1. Introduction

1.1. Plan of the review

The purpose of this review is to provide an introduction for physicists interested in understanding the basis for current concern that the 'greenhouse effect' may be changing the climate.

After some introductory definitions and discussion of the physics involved, the remainder of the review is in three main parts. In the first, we consider some specific elements of the whole, very complex, climate system, namely: the origin and basic structure of the atmosphere; its composition and photochemistry; natural biogenic and anthropogenic contributions to the budgets of atmospheric minor constituents; feedback processes, especially those involving hydrodynamics, clouds and aerosols; and the oceans, including their role as vast reservoirs and transporters of heat, water vapour and carbon dioxide for the atmosphere. Next, we see how the various processes can be brought together and the result of their interactions quantified by the use of physical models. We look at the prognoses provided by the most sophisticated currrent model studies, and the consensus on the range of uncertainties which remains. Then we consider how our understanding of physical processes and the limitations of models can be explored and uncertainties reduced by suitable measurements, and describe some of the leading programmes now underway.

In these discussions it is useful to consider the role of the greenhouse effect on the other terrestrial planets, where the same physics acting within different boundary conditions produces illuminating and often tantalising results.

1.2. What is meant by climate?

Climate is not a scientific term, and its meaning is not rigorously specified. Most definitions involve the long-term average of one of, or all of, a set of atmospheric variables, or a field of such values, usually measured at or near the Earth's surface. While it is quite reasonable, and is becoming increasingly common, to refer to the climate of a particular region of the upper atmosphere, or the deep ocean, or the surface of Mars, the term is especially appropriate when used in connection with the human environment and our associated comfort and well-being.

Different areas of research require different definitions of climate. To the physical geographer or meteorologist, 'long-term' usually means long enough for the day-to-day fluctuations in the temperature, say, or the rainfall, at a particular location to be smoothed out, and for this purpose a month is often taken to be a convenient period. He would probably consider behaviour which occurred consistently in monthly averages over a period of several years, or which occurred frequently during a decade, to be definitive in this context. Data for temperature, rainfall, and other meteorological parameters can be used to obtain monthly averages over five or ten years, or some other period, and these and the associated variances provide a quantitative definition of climate. In most traditional definitions the climate is regarded, within certain limits or 'natural fluctuations', implicitly treated as small, as invariant. If the mean value of a parameter changes progressively over a long period of time, and if this secular change exceeds the natural fluctuations, then the climate would seem to change; 'seem', because of course it is always possible that what seems like a trend is in fact a large amplitude variation of long period rather than a truly secular change.

In fact, we know that the Earth's climate has shown large variations in the distant past, from the evidence of the ice ages for example, and that it has fluctuated substantially in more recent times too. The more accurate records of the last century or so show changes of smaller amplitude on a variety of time scales. These fluctuations, usually discussed in terms of temperature as the most important and best-measured parameter, are of great interest because they may express the chaotic behaviour of the Earth-atmosphere-ocean-cryosphere system, or they may be the response to external forcing such as changes in the energy arriving from the Sun, or a combination of both. The system is so complicated and the historical record so incomplete that, at the present time, only modest progress has been made in understanding the mechanisms underlying natural variability and no physical model exists which can explain from first principles the fluctuations which appear in any kind of climate record.

In this review, we shall be considering the physical basis for a elementary understanding of climate, defined except where otherwise made clear as the mean temperature of the surface of the Earth, averaged over the seasons. We shall also discuss the reasons for the complexity of a more detailed understanding, such as will be required if long-term weather forecasting (inevitably, with low resolution in space and time, and so perhaps better called 'climate' forecasting) is to become a practical proposition. These are part of what might be called the 'traditional' climate problem.

Overshadowing such questions in the public eye at the present time, although related to them, is the possibility of persistent *anthropogenic* climate modifications, that is, quasi-permanent changes driven by human activities. We shall see that some change is the inevitable consequence of atmospheric pollution, although the same kinds of complexities which make natural climate change hard to predict, cloud the crystal ball when it comes to forecasting global change in detail. To improve matters, we need to understand the geosystem better, with more and better measurements and realistic, physics-based computer models two of the top priorities.

1.3. The energy balance of the Earth as a planet

The Earth's annual mean surface temperature is determined by the balance between the incoming energy from the Sun, mostly at wavelengths near that of visible light, and outgoing infrared energy from the surface and atmosphere. Both are modified by optical and thermodynamic processes in the atmosphere, involving in particular water in all three of its phases.

The effective (equivalent blackbody) temperature, T_S, of the surface of the Sun is about 6000 K, its radius R_S is about 0.7 million kilometres and its distance from the Earth D_{ES} is about 150 million kilometres. Applying the Stefan–Boltzmann law we obtain for the total radiant power of the Sun:

$$E_{Sun} = 4\pi\sigma R_S^2 (T_S)^4 \text{ W} \tag{1}$$

where the constant σ is equal to 5.670×10^{-8} W m^{-2} K^{-4}. The solar constant S, which is defined as the power per unit area arriving at the Earth from the Sun, is then given by

$$S = \sigma T_S^4 (R_S/D_{ES})^2 \text{ W m}^{-2} \tag{2}$$

from which we find that S is approximately 1.37 kW m^{-2}. In order for the planet to be in equilibrium overall, the power emitted by the Earth over its whole area, E_E, must be equal to the total incoming power of the Sun which is intercepted by the projected area of the Earth, less the fraction A which is reflected, i.e.

$$E_E = 4\pi\sigma R_E^2 (T_E)^4 = (1-A) S \pi R_E^2 \text{ W} \tag{3}$$

from which we obtain the result that T_E is approximately 250 K, or −23 °C. This must be the average equivalent blackbody temperature of a solid sphere the same size and emissivity as the Earth at the same distance from the Sun, but as a value for Earth's mean surface temperature it is clearly wrong. The reason for the discrepancy is of course the atmosphere, which lies over the surface of the Earth like a blanket. The equilibrium temperature which we have just calculated is the temperature of the outside of the blanket; the inside is warmer.

The energy from the Sun reaches the surface of the planet because solar radiation consists mostly of photons at visible wavelengths and the atmospheric blanket is nearly transparent at those wavelengths. The balancing, cooling radiation, however, is in the infrared part of the spectrum, because the source is so much cooler, a few hundred degrees instead of a few thousand (figure 1).

Wein's law states that the peak emission occurs at a wavenumber equal to approximately twice (actually 1.962 times) the temperature of the source, i.e. 500 cm^{-1} or 20 μm for the Earth and 12 000 cm^{-1} or 0.8 μm for the Sun. At infrared wavelengths,

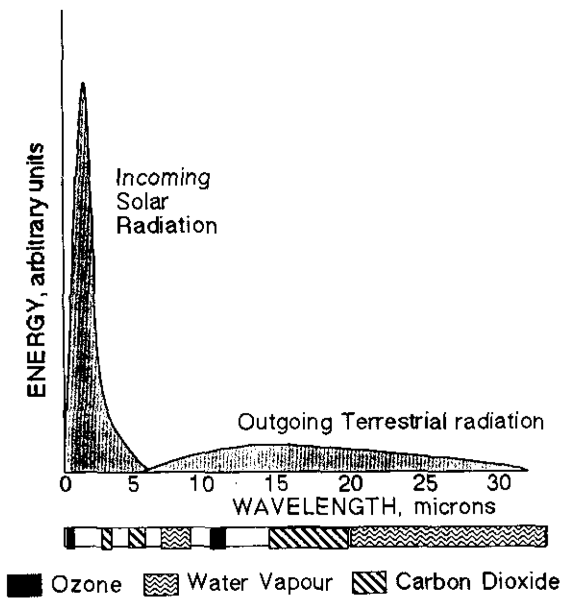

Figure 1. Emitted energy wavelength for blackbodies at temperatures of 6000 K and 250 K, approximating the Sun and the Earth respectively. The approximate locations in the spectrum of the strongest bands of the three most important minor constituents are shown at the bottom.

many of the minor constituents of the atmosphere absorb strongly in their vibration-rotation bands. (Interestingly, the major constituents, nitrogen and oxygen, as homopolar molecules without dipole moments, contribute only weak pressure-induced absorption bands and so play no major part.) Thus the blanket is effective against outgoing but much less so against incoming radiant energy, producing a net warming which overall for the Earth is about 33K (figure 2).

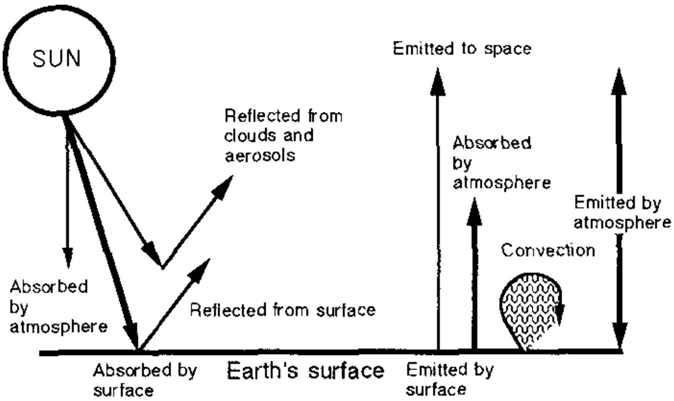

Figure 2. A schematic diagram illustrating the main components of the greenhouse effect.

The name 'greenhouse' for this atmospheric effect arises because the glass in a greenhouse is transparent to sunlight, but not to infrared radiation, in much the same way (although inhibition of convection makes the more important contribution to the heating of a domestic greenhouse). 'Blanket' effect might be a more accurate and more graphic name to use, but in either case it is crucial to remember when using commonplace examples that the atmospheric blanket (or greenhouse) is not a simple thing. For one thing, its thickness is more wavelength-dependent in reality than in the simple visible-transparent, infrared-opaque representation, and for another, its detailed make-up involves motions, phase changes and other thermodynamics, so models based on radiative balance alone can be misleading. It has been calculated, for instance, that an atmosphere in purely radiative equilibrium (i.e. an imaginary one which is not allowed to move), would support a mean global surface temperature of 72 °C (Moller and Manabe 1961).

Vertical convection of sensible and latent heat accounts for the difference between a radiative balance calculation of the surface temperature, and reality. We need to understand what changes to expect, in response to a tendency towards global warming forced by increased concentrations of greenhouse gases, in these processes too before the net outcome can be predicted with confidence. Some respected scientists (e.g. Lindzen 1990) think that the Earth's hydrosphere may be a natural thermostat which could mitigate enhanced greenhouse warming to the extent that the issue loses its urgency on the geopolitical scene. Others dismisss this as metaphysical (Hansen *et al* 1990).

1.4. Greenhouse gases and their changes

The Earth's greenhouse is maintained by a cocktail of minor constituents in which H_2O and CO_2 dominate but CH_4, N_2O, CFCs (chlorofluorocarbons) and other trace gases, as well as clouds and aerosols, contribute significantly. Different gases contribute to greenhouse warming to a different degree. The contribution of a given species depends not only on how much of it is present, but also on the number and probabilities of the various transitions responsible for its infrared spectrum, the location of the vibration–rotation bands relative to the maximum of the Planck function, and the degree of overlap with the bands of other species. The number of bands tends to increase with the number of atoms in the molecule, as more modes of vibration are available in a more complex molecule. Diversity is important because spectral bands saturate, so that adding more of a gas has a relatively small effect on those of the strong bands which are already strongly absorbing, as for the strong fundamental band of CO_2 near 15 μm. The additional opacity can be thought of in terms of increasing only the spectral 'width' of the band, and not its depth. For a single isolated spectral line, the effect on the optical depth of the atmosphere (defined as the log of the transmission over a specified path) is proportional to the mass of gas per unit area in the path if the absorption is weak, i.e. if the transmission is non-zero in the line centre. As the concentration of absorber increases, the line saturates, the line centre blacks out, and the optical depth gradually changes to a dependence on the square root of absorber amount (e.g. Goody and Yung 1989). It follows that adding a quantity of a gas present initially only in extremely small amounts (like the CFCs) will tend to have a larger effect than adding the same amount of a relatively abundant minor constituent like carbon dioxide, provided of course that the former has spectral bands which are not obliterated by overlapping bands of strongly absorbing gases like CO_2.

The most common CFCs, 11 and 12, have bands in the 8–14 μm atmospheric window, and hence are relatively efficient greenhouse enhancers if their abundance increases. This efficiency can be quantified by defining the 'global warming potential' of a gas as the increase in surface temperature produced by unit mass of the gas, compared with the effect of the same amount of CO_2. Estimates (from models) of this quantity for various gases are shown in table 1. Also shown are the current rates of emission of these gases into the atmosphere, and model estimates of how much each will contribute to global warming in the next 100 years.

We see from table 1 that CO_2 is the most important pollutant for greenhouse purposes, not because of its efficiency (the strong CO_2 bands are already saturated, as

Table 1. The principal greenhouse gases with substantial man-made components, their present mean atmospheric concentrations (In ppmv, or parts per million by volume), their 1990 emission rates, their relative efficiencies for global warming (see text) and their estimated relative contributions to recent warming. After IPCC (1990) and Deutscher Bundestag (1989).

	Present concentration (ppmv)	Present emission rate (teragrams per year)	Global warming potential	Relative warming (%)
CO_2	346	26 000	1	60
CH_4	1.65	300	21	15
N_2O	0.31	6	290	4
CFC 11 & 12	0.0005	0.9	5800	11
HCFC 22	0	0.1	1500	0.5
Ozone	$\leqslant 10$	—	2000	8
Other				0.5

discussed above) but because there is so much of it being produced, and because its second-strongest band, the bending fundamental at 15 μm, falls right at the peak of the Planck function for typical terrestrial temperatures. (The 'secondary' greenhouse agents, water vapour and cloud, which do not depend on human activities, are not included in the table, but their feedback roles are estimated in the models).

All of the greenhouse gases listed above are produced by human activities, and increasing industrialisation has meant that the balance between production and removal has been shifted in favour of higher atmospheric concentrations. It is fairly obvious even before performing any calculations that these will tend to increase the greenhouse effect. The more difficult question that we will come to below is to estimate the rate at which the increase will occur.

One of the major unknowns is what the concentrations of the important minor and trace gas concentrations will be in the years ahead, since we can only guess at the likely rates of production more than five or ten years ahead. Also, the removal processes, even for a relatively simple case like CO_2, are not known quantitatively in detail either. To embrace these uncertainties, modellers have introduced 'scenarios', in which the effects of extrapolating recent trends are estimated along with higher estimates (such as may occur if, for instance, industrialization of the developing nations proceeds faster than expected or natural removal is slower than expected) and lower estimates (for situations like those where controls on industrial emissions are successfully adopted). Such scenarios for the principal man-made greenhouse gases are discussed in section 2.3.2.

CO_2 has increased by 25% in the last 100 years, and continues to rise at a rate of 0.5% p.a.; methane concentrations have doubled in the same period and are increasing at 1% p.a.

The mean level of water vapour is in a kind of dynamic equilibrium with the liquid and solid water in the oceans and cryosphere; any increase in mean surface temperature due to increases in other gases such as CO_2 will tend to be amplified by the increased capacity of the lower atmosphere for water vapour. H_2O is in fact the most important greenhouse gas in terms of its contribution in degrees to the surface temperature, since it is present in enormous quantities. However, anthropogenic release of water vapour is not itself a direct contributor to enhanced greenhouse warming. The role of water vapour is therefore one of amplifying changes forced by other gases. In most important studies using models (see e.g. IPCC, 1990) the amplification due to water vapour of the increase due to doubling CO_2 is about 60%. Others (e.g. Lindzen 1990) suggest that it could be negative if changes in the tropospheric circulation lead to a drying of the upper troposphere.

1.5. Clouds, aerosols and associated uncertainties

Some scientists active in the field of climate research remain sceptical about the possibility of global change resulting from an enhanced greenhouse effect, and even those in the majority who do believe global warming to be a threat are always careful to draw attention to the substantial uncertainties involved in making any forecast of a fundamentally chaotic system. It is in the area of the physics of clouds and aerosols that the most serious difficulties arise. One example of disagreement over even the sign of one important effect is quoted at the end of the last section.

High clouds (altitudes above 6 km or so) are often made of ice crystals, which scatter solar radiation strongly in the forward direction, i.e. towards the surface. At the same time, they can present a substantial optical depth to upwelling long wave radiation. The combination therefore tends to warm the troposphere. Low clouds, on the other hand, are more reflective than the surface below on average, and so tend to increase the reflectivity or albedo of the planet, resulting in a net cooling.

The exact behaviour of a given cloud depends on its height, water content, particle size distribution and other factors sometimes collected under the term 'microphysics'. Calculating the behaviour of a single cloud, even given a knowledge of these variables, is just about within the state of the art, provided the cloud is not too inhomogeneous. There is little early prospect of a precise solution to the net effect on the entire globe of the myriad of clouds of all types which are scattered around the world at any given time; theorists have to resort to simplified representations of cloud distribution and behaviour (termed *models* and *parametrizations* respectively). Getting this right is a formidable task even when modelling a past state of the atmosphere for which extensive observations may be available. One must then ask how the cloud field is going to evolve *in the future.* At present, we cannot be sure if a future world with an enhanced greenhouse effect will be more or less cloudy than now, let alone what the precise radiative properties of the new distribution will be.

Clouds are not the only problem. Hansen *et al* (1990) have stressed the importance of aerosols other than cloud droplets themselves, i.e. the background turbidity of the atmosphere. They believe that this is the single greatest uncertainty we face about climate forcing. Aerosols exist in the stratosphere, most notably as very small droplets of sulphuric acid, occurring as a result of natural processes (principally volcanic eruptions) but also of anthropogenic processes (factory and car emissions). Tropospheric aerosols include water and acid droplets, dust and soot particles, and salt crystals. Their amounts, trends and variability are very little known, and their effect on the climate is obscure but probably quite significant. Aerosols, such as stratospheric sulphuric acid droplets, tend to cool the planet by backscattering sunlight efficiently, and there is some evidence that their concentration is growing. The notion has even been advanced that any restrictions on the burning of fossil fuels may *accelerate* global warming, because the aerosol effect, enhanced by sulphurous emissions, may more than offset the effects of increasing CO_2. This kind of argument was the basis for the concerns articulated during the 1970s that the Earth might actually be forced by global pollution into another ice age. This possibility, although not now thought probable, still cannot be entirely discounted. It illustrates the difficulties climate modellers will continue to have in discussions with policy makers and the media, until the spread of uncertainty in the models can be further reduced.

1.6. Climate and climate change on the terrestrial planets

The 'blanket' analogy introduced in section 1.3 is better for Venus than for the Earth. The sulphuric acid droplets in the Venusian clouds are highly reflective at short wavelengths and so scatter sunlight and diffuse it in all directions with a high efficiency. This gives Venus its high albedo (see below), but also allows a few percent of the incident solar energy to penetrate the thick cloud layers and reach the surface. The same cloud droplets are quite black at longer wavelengths, and so they strongly absorb upwelling long-wave radiation at the same time—a paradigm of greenhouse behaviour. The Earth's atmosphere, on the other hand, is relatively thin, and the blanket has spectral gaps in the so-called 'window' regions, where strong absorption bands are absent, the opacity is small, and the surface can cool directly to space (figure 2). On

Venus, the windows are mostly blocked well above the surface by the large amounts of carbon dioxide, water vapour and other infrared-active molecules below the clouds, as well as the clouds themselves.

The calculation of an equilibrium blackbody temperature T_V for Venus produces a number which is lower than the corresponding value for the Earth, in spite of the greater proximity to the Sun, because Venus has such a high albedo, reflecting 76% of the incoming radiant energy. The value for T_V obtained from (2) is 240 K, which is in fact the temperature of the top of the thick, unbroken cloud layer which covers Venus. The surface is at a temperature of around 730 K. The greenhouse enhancement on Venus is therefore about 490 K. How this large warming evolved, whether it subject to change at present, and what it means for the Earth is an important research topic (Hunten 1989). The general message that earth-like planets clearly can be subject to extremely large greenhouse warmings is an important and often underrated one.

It is well known that the climate of the Earth has changed very substantially in the past, because of the geological and other evidence of ice ages. Although the underlying mechanisms remain at least partially obscure, a changing greenhouse mechanism seems to be at least part of the picture. Data from the analysis of Arctic ice cores shows large and strongly correlated changes in temperature and CO_2 abundance over the last 100 000 years (IPCC, 1990). Unfortunately, however, the resolution of the data along the time axis is not good enough to say which quantity leads the other, and of course we do not know what else may have been changing at the time. In principle, increases in both temperature and CO_2 could be driven by a changes in a third variable, the solar constant for example.

On Mars, we observe a significant greenhouse effect driven mainly by dust (section 4.3) and see definite evidence of a much warmer, wetter climate in the distant past. The powerful Venusian greenhouse, which, as we have seen above, produces nearly 500 K in surface temperature enhancement, is possibly not static, although we have no way to know from the sparse data available. Nor is our theoretical understanding of the atmosphere of Earth's neighbour and near-twin good enough for us to say at present whether the extreme greenhouse conditions appear stable on theoretical grounds. Nevertheless, the existence of a huge greenhouse effect on Venus and the evidence for change on Mars both drive home the point that the Earth's greenhouse could be in a state of delicate balance. In the case of the Earth, the best calculations seem to support this inference, although we should keep in mind that scenarios can be developed which support those who think that the Earth has inherent stabilizing mechanisms which tend to keep the climate 'locked in' on the best conditions for life.

Notwithstanding the dissenters and the uncertainty over the details, the threat of significant, short-term climatic change in response to changes in atmospheric composition which are certainly taking place at the present time is very real. We should like to have the means to predict the time scales on which changes of a given magnitude will occur, and also to understand the details of the change, for example its seasonal or geographical dependences. This is largely a problem in physics, the basis for and current status of which is discussed in the sections below.

2. Components of the climate system

In this section, we consider some of the fundamental properties of the major components of the climate system, beginning with the origin and basic structure of the atmosphere. In considering the crucial topic of atmospheric composition, it is necessary to take account of photochemical processes which modify the composition of the atmosphere in important ways, including the production of ozone; the role of the biosphere, including the anthropogenic production of greenhouse gases; feedback processes involving water vapour and clouds; and the interaction of the oceans with the atmosphere.

2.1. Atmospheric structure

2.1.1. Origin of the atmosphere. The Earth's atmosphere is believed to be of secondary origin. This means that the gases surrounding the solid planet at the time it condensed were lost to space at an early stage and the present atmosphere subsequently formed as a result of outgassing from the crust. This process was augmented to an unknown degree by the infall of icy cometary material. The primitive atmosphere thus formed would have had a composition quite different from that which exists today, consisting primarily of carbon dioxide, methane, ammonia and water vapour with no free nitrogen or oxygen. The large proportion of nitrogen now present arose as a consequence of the photodissociation of ammonia into nitrogen and hydrogen. The latter gas is so light that it can escape from the Earth even at the present time, whereas most heavier gases are gravitationally trapped now that the Earth is cooler and the Sun less active. As the atmosphere, and life, evolved, most of the methane was oxidized to form carbon dioxide and water vapour, although enough is produced (for example, by the decay of vegetable matter) to leave a substantial admixture in the atmosphere to the present day. Free oxygen exists as a consequence of the presence of life, in the form of plants which convert carbon dioxide into oxygen by photosynthesis. Thus, we owe the present state of the atmosphere to a series of complex and dynamic processes. For a longer review see, for example, Levine (1985).

2.1.2. Vertical structure. Figure 3 shows a simplified model of the various regions of the atmosphere, which are named for their temperature structure. As noted above, a substantial fraction of the energy from the Sun reaches the ground where it is absorbed. The lower atmosphere (troposphere) is then heated from below by the ground and becomes convectively unstable. Large-scale overturning, coupled with some radiative and turbulent transfer, advects energy upwards until it reaches a level where the overlying atmosphere is of a sufficiently low optical thickness that significant amounts of radiative coooling to space in the thermal infrared can occur. At this level, the *tropopause*, convection ceases and the temperature tends to become constant with height. The tropopause occurs at a level of about 10 to 16 km above the surface, depending mainly on latitude, and is generally quite a sharp feature in the temperature profile.

The temperature gradient in the convectively mixed region can be calculated, as follows. Assuming that hydrostatic equilibrium applies, pressure P is related to density ρ at a given height z by

$$\mathrm{d}P = -g\rho\, \mathrm{d}z. \tag{4}$$

If we assume that there is no net exchange of energy between a parcel of air and its surroundings, then specific heat at constant pressure c_p, temperature T, pressure P, and volume V are related by

$$c_v\, \mathrm{d}T + p\, \mathrm{d}V = 0. \tag{5}$$

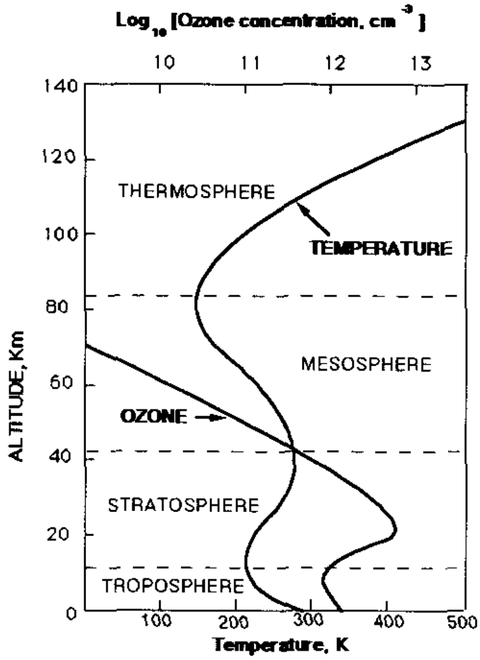

Figure 3. The mean temperature structure of the atmosphere, showing the names given to various regions.

Using the perfect gas law $PV = RT/M$ (R = gas constant per mole, M = molecular weight), and the relationship $c_p - c_v = R/M$ (c_v = specific heat at constant volume) we find that the temperature gradient with height $\mathrm{d}T/\mathrm{d}z$ is given by

$$\mathrm{d}T/\mathrm{d}z = -g/c_p. \tag{6}$$

This gradient, called the lapse rate, is constant, and, since c_p is equal to about $1000\,\mathrm{J\,kg^{-1}\,K^{-1}}$, equal to approximately $10\,\mathrm{K\,km^{-1}}$ for dry air. For moist air, $\mathrm{d}T/\mathrm{d}z$ is less and can be as small as $3\,\mathrm{K\,km^{-1}}$. $6.5\,\mathrm{K\,km^{-1}}$ is a useful average value for Earth.

The lapse rate above the tropopause, where convection stops, tends to zero (i.e. constant temperature with height) because there is no longer enough absorption above the layer to stop emitted photons reaching space. Then each layer is heated by radiation from the optically thick atmosphere below, and cooled by radiating to space, to the same degree; to first order height is no longer important. This region is called the stratosphere, since it is stratified in the sense that the layers are not convectively unstable as they are in the troposphere.

The stratospheric temperature T_x may be estimated by treating the region as if it were a single slab of gas which is optically thin at all wavelengths, rather than just on average, which is the real situation. ('Optically thin' means that a photon traversing the slab will do so with a probability of not more than $1/e$ of being absorbed or scattered). Then T_x is related to the effective radiative temperature of the Earth T_E (as defined and evaluated in section 1.3) by the expression for the energy balance of the stratosphere, treated as a slab of emissivity and absorptivity ε:

$$\varepsilon\sigma(T_\mathrm{E})^4 = 2\varepsilon\sigma(T_x)^4 \tag{7}$$

whence

$$T_x = \frac{T_E}{2^{1/4}} = \frac{250}{2^{1/4}} \approx 210.2\ \mathrm{K}. \tag{8}$$

Although these simple arguments suggest that the stratospheric temperature remains constant to a great height, in fact the value for T_x just calculated applies only from the tropopause up to about 20 km. Above this, temperature is observed to increase (figure 3) to a maximum value of around 270 K near 50 km altitude. This is a consequence of the absorption of solar ultraviolet radiation by the stratospheric ozone layer (section 3.3).

The height of the temperature maximum is known as the *stratopause.* Above the stratopause, the temperature declines again, reaching an absolute minimum of about 190 K at the *mesopause* near 85 km, where the pressure is only a few microbars (1 bar = 10^5 Pa is a standard surface pressure). Energetic particles and solar photons in the extreme ultraviolet penetrate in to the region above the mesopause, causing ionization and dissociation of air molecules and releasing kinetic energy. The heating thus produced causes the temperature to increase rapidly with height, leading to the name *thermosphere.*

A fairly small distance up into the thermosphere, at around the 100 km altitude level, the atmosphere ceases to be mixed by turbulence and starts to separate into its lighter and heavier components. For present purposes, this level (the *homopause*) may be considered to be the effective top of the atmosphere.

2.2. Composition and photochemistry

2.2.1. Composition of the troposphere. Table 2 shows the typical composition of dry air near the surface. The amount of water which is present in addition to these non-condensible species varies, but can be as high as a few per cent. Some of the other species (e.g. ozone, carbon monoxide) are also very variable even in non-polluted environments.

2.2.2. Tropospheric and stratospheric ozone. Tropospheric ozone is an important pollutant because of its toxic effects, its role in atmospheric chemistry involving other

Table 2. Composition of clean dry air near the surface.

Species	Mixing ratio (by volume)
Nitrogen, N_2	0.780 83
Oxygen, O_2	0.209 47
Argon, Ar	0.009 34
Carbon dioxide, CO_2	0.000 35
Neon, Ne	18.2×10^{-6}
Helium, He	5.2×10^{-6}
Methane, CH_4	2×10^{-6}
Krypton, Kr	1.1×10^{-6}
Hydrogen, H_2	0.5×10^{-6}
Ozone, O_3	0.4×10^{-6}
Nitrous oxide, N_2O	0.3×10^{-6}
Xenon, Xe	0.1×10^{-6}
Carbon monoxide, CO	0.1×10^{-6}

species, and its infrared opacity in the 9–10 μm wavelength region, which contributes significantly to the atmospheric greenhouse. Although its overall budget is poorly understood, tropospheric ozone originates both on and near the surface and in downward propogation from the stratosphere where the concentration is higher. Tropospheric ozone is generally on the increase, in contrast to the stratospheric ozone layer for which there are much-publicised fears of a trend towards lower concentrations.

Stratospheric ozone depletion and global warming due to the greenhouse effect are separate but related problems, the main connection being that the ozone layer has a small but not insignificant effect on the flux of solar energy reaching the surface, and a small but again significant contribution to the opacity of the atmosphere in the thermal infrared. Both are cases where the natural state is being perturbed by increasing amounts of man-made materials. The importance of the ozone layer in determining the temperature structure has been mentioned above (section 1.2), and its contribution to global habitability, by reducing the amount of energetic and potentially damaging ultraviolet radiation which reaches the surface, is well known. Ozone is a very unstable molecule, and the amount which is present in a given parcel of air at a given time depends on a balance between production and loss mechanisms. In their simplest form, as first enunciated by Chapman in the 1920s, the equations may be written

$$\begin{aligned} &O_2 + h\nu = O + O \\ &O_2 + O + M = O_3 + M \end{aligned} \tag{9}$$

for production, and

$$\begin{aligned} &O_3 + h\nu = O + O_2 \\ &O + O_3 = O_2 + O_2 \\ &O + O + M = O_2 + M \end{aligned} \tag{10}$$

for destruction. To obtain reasonable agreement with measurements of the ozone abundance as a function of height, the Chapman scheme has to be augmented to include the effects of catalytic reactions involving trace constituents such as oxides of nitrogen and chlorine. Even then, the amount of ozone varies substantially, not only with latitude and season (as a consequence of the changing solar zenith angle, and of variations in the circulation of the atmosphere) but also in an apparently random fashion. Research driven by the current concern about ozone depletion (section 8.1) is made complicated by the existence of these large natural fluctuations, many due to causes which are as yet imperfectly understood.

2.2.3. Nitrogen cycles. In addition to the molecular nitrogen which makes up most of the atmosphere, there is a small amount of nitrogen present as the oxides and as nitric acid. The main source molecule is N_2O, which originates at the surface from various sources including nitrogen-fixing bacteria and artificial fertilizers in the soil. Once in the stratosphere, photodissociation of N_2O produces NO which reacts with ozone:

$$\begin{aligned} &NO + O_3 = NO_2 + O_2 \\ &\underline{NO_2 + O = NO + O_2} \\ &\quad O + O_3 = 2O_2 \end{aligned} \tag{11}$$

In this reaction ozone is destroyed but the nitric oxide is regenerated. This catalytic cycle is one of the most important of many which affect the balance between ozone creation and destruction and hence tend to reduce the mean amount present in the stratosphere.

The time constant for a molecule of N_2O, newly released at the surface, to become involved in stratospheric ozone chemistry is long, of the order of 10 years. The time constant for its eventual removal from the stratosphere (probably by re-entering the troposphere, perhaps as nitric acid, and then by rainout onto the surface) is even longer, of the order of 100 years. Thus, the role played by this and other species in ozone depletion and in greenhouse enhancement, is slow to build up but cumulative. Present conditions reflect the emissions of past decades and any trends which may be present now will continue to grow for some time to come, even if harmful emissions are stabilized.

2.2.4. Chlorine cycles. Small amounts of free chlorine are also present in the stratosphere, following the photolysis of chlorine-containing compounds, mainly man-made chlorofluorocarbons, especially CFC-11 and CFC-12 (the notation refers to the chemical formula, where CFC-nm is $C_nF_mCl_{4-m}$). These species are very inert under room conditions and are widely used in industry as refrigerants, propellants, and in various other ways. In the stratosphere, however, they are decomposed by photolysis or by reaction with OH radicals:

$$CH_3Cl + h\nu = CH_3 + Cl$$

$$CH_3Cl + OH = CH_2Cl + H_2O.$$

The chlorine cycles involving ozone are complex, but basically consist of the catalytic cycle

$$\begin{array}{c} Cl + O_3 = ClO + O_2 \\ \underline{ClO + O = Cl + O_2} \\ O + O_3 = 2O_2 \end{array}. \tag{12}$$

At the present time, the nitrogen cycle dominates that of chlorine for ozone depletion but stratospheric chlorine concentrations are increasing by about 5% per annum at the present time and represent a potentially serious threat for the future. It has been calculated that failure to curb CFC production will remove 10% of the vertical ozone column at mid-latitudes by the year 2050.

2.3. Biosphere

The two principal ways in which life-forms on the Earth modify the greenhouse are the roles of plant life on land and in the sea in determining the carbon dioxide abundance in the atmosphere, and the role of humans in producing CO_2 and the other greenhouse gases as a by-product of their burgeoning industrial activities.

2.3.1. The carbon cycle. Figure 4 shows schematically the principal reservoirs of carbon in various forms which reside within the geosystem and the fluxes between them which determine the amount of CO_2 in the atmosphere at any given time. The principal man-made fluxes, due to fossil fuel burning and deforestation, are small compared to the natural exchange between plants, the soil and the oceans. Thus, it is necessary to be concerned not only about man-made production of CO_2 itself, but also any other pollution which may affect the natural equilibrium, for example by reducing the

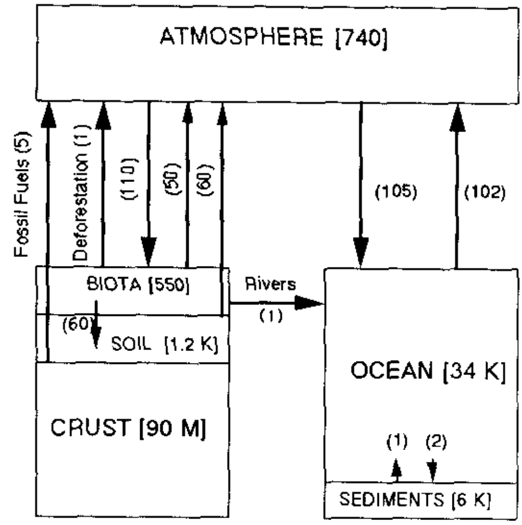

Figure 4. The global carbon cycle. Estimates of the principal reservoirs are shown in units of 10^{12} kg of carbon, and fluxes in units of 10^{12} kg for year.

population of biota in the oceans. On the other hand, there is a possibility that the take-up of CO_2 by the oceans may *increase* if warmer water leads to an increase in aquatic plant life, in spite of a reduction in the solubility of the gas in warmer water. More vigorous plant growth on land, as a result of higher atmospheric CO_2 concentrations, may also tend to slow down the build-up of the gas in the atmosphere.

2.3.2. Mankind's influence on atmospheric composition. Water vapour is the most effective component of the unperturbed greenhouse, contributing about 65% of the 33 K of 'natural' warming which the planet enjoys. It is nevertheless true that greenhouse change is being forced by the man-made additions to the other infrared active minor components of the atmosphere. The principal participants, their emission rates and their relative contributions to incremental global warming are listed above in table 1.

While most emission rates can be estimated reasonably reliably, the actual rate of build-up of greenhouse gases in the global atmosphere is much harder to determine, because it requires estimates of the loss rates due to processes such as rainout, sedimentation, photolysis, photosynthesis which are difficult to quantify. For species like the CFCs, lifetimes are on the order of centuries and the gases (so far as is known) accumulate essentially indefinitely. The lifetime of methane is a few years, and the product is another important greenhouse gas-stratospheric water vapour. (Because of the cold-trapping effect of the tropopause, methane oxidation at least as important as transport from the troposphere as a source of stratospheric H_2O.) The budget of CO_2 involves the oceans and plant life as described elsewhere (sections 2.3.1 and 2.3.3).

Estimates of the future build-up of the key greenhouse gases have been made, in spite of the difficulties in so doing, since they are essential to any attempt to make a quantitative forecast of future global warming. Figure 5 shows the recent and predicted rates of change in the principal man-made greenhouse gases from one source (Deutscher Bundestag 1989). The uncertainty in the prediction increases with time into the future, partly for the reasons stated above, and partly because of uncertainties due to political interventions like the Montreal Protocol on CFC emission. The middle curves in each

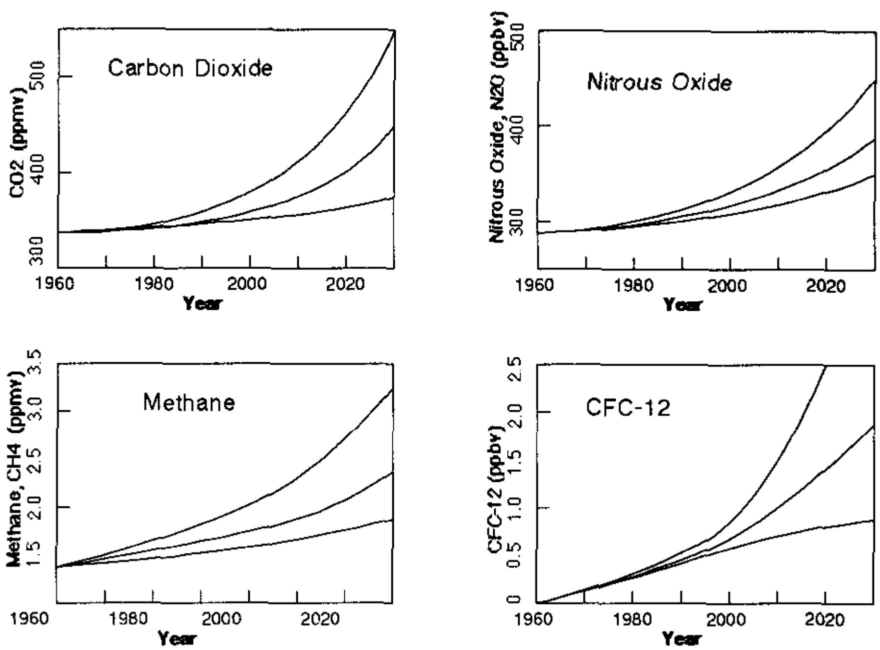

Figure 5. One estimate, with uncertainties, of the concentrations of the principal greenhouse gases as a function of time for carbon dioxide, methane, nitrous oxide and chlorofluorocarbon 12. After Deutscher Bundestag (1989).

case assume that the emission rates remain at their present level, which is fairly optimistic. Predictions of the global warming expected to result from such changes are discussed later.

2.4. *Feedback processes*

Some feedback processes in the global carbon cycle, some tending to accelerate and others to slow the buildup of atmospheric CO_2, have already been mentioned (section 2.3.1). There are many others which are understood qualitatively but which are very difficult to incorporate accurately into models, principally the following.

2.4.1. Water vapour. As the lower atmosphere warms due to the addition of more anthropogenic greenhouses gases like CO_2, it can hold more water vapour, which of course is freely available by evaporation from the surface. Thus, water vapour tends to amplify any increase due to other gases and calculations indicate that this amplification factor may be as much as a factor of two.

2.4.2. Methane. A smaller, but still important, amplification factor may be due to methane, the production of which in marshland, etc, is temperature dependent. It has been postulated that very substantial additional amounts of methane may be released from subarctic soils and tundra as they thaw, thus amplifying any warming trend.

2.4.3. Clouds. The enhancement of the greenhouse by increased tropospheric humidity may be offset, or under certain circumstances amplified, by corresponding changes in the cloud cover (section 1.3). The infrared properties of clouds, as well as cloud height and cover and the abundance of water vapour, are highly variable quantities and, with aerosols, represent the greatest source of uncertainty in the understanding of the climate system. The main effect of low clouds is to increase the albedo of the Earth (section 3.3.1) and hence to cool the whole troposphere including the surface. High clouds, however, are made of large ice crystals and are generally strongly forward-scattering at short (solar) wavelengths, while having large absorption coefficients for upwelling long-wave radiation. These clouds could increase the greenhouse effect substantially, even in modest amounts, such as those which could be produced by fleets of high-flying aircraft.

The atmosphere contains numerous other cloud types and thinner layers of aerosols which may have important roles. Tropospheric aerosols are necessary for cloud nucleation, and changes in the number or type of aerosols could be crucial in determining how the Earth's cloud cover evolves in the future. In the stratosphere, the Junge layer, which consists of small sulphuric acid droplets, is thought to be increasing in density due to increasing sulphur emissions added to volcanic activity. In some analyses, this will have an important effect by increasing the albedo of the planet.

The understanding of cloud physics which is needed to tackle their role in the present and future climate system is daunting, especially when compared to the present state of knowledge. There are many different classifications of clouds, and the details of their formation and evolution depend on the state of the rest of the climate system itself (Arking 1990). However, nearly all form by some version of the same basic process: convection upwards of moist air and subsequent condensation of water droplets or ice crystals around dust nuclei (Mason 1974). A typical cloud droplet is about 10 μm in diameter, while a typical raindrop is one hundred times larger, i.e. contains as much water as a million cloud particles. Cloud droplets grow by diffusion and condensation of water vapour until they reach about 30 μm diameter, then by collision and coalescence until they eventually rain out.

For computing the radiative properties of clouds, in particular their transmissivity and reflectivity over the whole range of wavelengths important to the energy balance of the Earth (which as we have seen, e.g. from figure 1, is about 0.2–100 μm), the most important parameters are the physical extent of the cloud in three dimensions (x and y in the horizontal, and z in the vertical), and the size distribution $n(r, x, y, z)$ where $n\,\mathrm{d}r$ is the number of droplets of radius r to $r+\mathrm{d}r$ in unit volume of space with coordinates (x, y, z). In practice, r is usually assumed to be independent of x and y, and sometimes of z also; the simplest scattering calculations replace $n(r)$ with a single effective or 'mode' radius $\langle a\rangle$. Ice clouds present a particular problem, since the particles are crystals, often shaped like plates or needles or irregular forms with no spherical symmetry. Including the shape and orientation of the crystals is such a daunting part of the problem of computing their absorption and scattering properties that the modeller is reduced to introducing an equivalent spherical mode radius in this case also, citing probable random orientation of the particles as justification for doing so.

Then, if the refractive index of the cloud material is known, the cross-section for absorption and scattering, and the phase function which describes the directional distribution of the scattered photons, may be obtained for a single droplet from Mie theory (e.g. Goody and Yung 1989). Several algorithms exist (Hansen and Travis 1974) which allow the contributions of an assemblage of droplets to be built up into functions or matrices which describe the behaviour of a complete cloud. Even with the various simplifications mentioned, the computer power required for radiative transfer calculations in clouds is formidable, and the climate modeller generally resorts to relatively crude parametrizations which by-pass the physics altogether. Probably the best hope for the immediate future is that cloud measurements from satellites and aircraft (section 4), combined with theoretical case studies, will suggest improved parametrizations which represent the behaviour patterns of real clouds more reliably than is possible at present.

2.5. Atmosphere–ocean coupling

The oceans are vast reservoirs of heat, momentum, moisture, and dissolved gases (including carbon dioxide), all critical components of the climate system, and all of them quite efficiently exchanged between ocean and atmosphere. Consider, for example, that the top five metres of the ocean stores more heat than the whole of the atmosphere. This energy is transported globally by ocean currents (driven by winds) and released, with major effects on regional climates. The general circulations of the atmosphere and oceans are so intimately interconnected that most leading efforts to construct sophisticated climate models include the dynamics of both systems, and their coupling, in spite of the extra strain such a step imposes on computing resources.

By virtue of their long time constants for releasing heat energy, and for dissolving and releasing greenhouse gases, especially carbon dioxide, the oceans act to some extent as a 'thermostat' regulating, or at least slowing down, any global warming or cooling trend. At present it is estimated that about half of the annual human production of CO_2 (about 10 billion metric tons) is absorbed by the oceans, by dissolution, by chemical combination or by biological processes. Its ultimate fate is as estimated in figure 4, although the numbers are very uncertain. As noted above, this flux is actually a small net imbalance, at around the few percent level, in the much larger fluxes of CO_2 in both directions which occur across the atmosphere–ocean boundary each year.

A crucial factor in properly understanding and modelling the oceans' role in climate change is the relative difficulty in obtaining sufficiently comprehensive measurements. Efforts in this direction are examined below (section 4.1.3).

3. Models of climate change

Studies of greenhouse change, like virtually all problems involving complex coupled systems and a multitude of different processes, rely on models. Simple models provide useful insight into the physics involved, while complex models are necessary to obtain the greatest possible accuracy and to attempt to incorporate all of the feedback process. A major goal of modellers is to predict the future evolution of the climate system with some reliability. At the moment, this is a remote goal except for very general predictions. For example, from model studies it can be said with a high probability that global temperatures will rise, and global stratospheric ozone amounts decline, by amounts of a few degrees and several per cent respectively in the next few decades. Models are capable in principle of providing much more detail than this, but the current state of the art falls well short. At the present time, modellers are still striving to produce fully accurate representations of the *present* climate from first principles, and there is still no consensus about which aspects of the apparently chaotic climate system predictions are possible on any particular time scale.

3.1. Simple physical greenhouse models

The purpose of this section is to illustrate how basic physics and semi-empirical parametrizations can be used to construct simple climate models.

3.1.1. An elementary model. A very simple model, with some realistic features, can be constructed as follows. Let the atmosphere be represented by a single homogeneous layer of gas of temperature T_a. As noted above (section 1.3), to first order the gases of the atmosphere are transparent at those wavelengths corresponding to most of the incoming solar energy, and opaque at those wavelengths at which the Earth emits most of its thermal energy. This behaviour can be approximated by assigning to the model atmosphere a bimodal transmission function equal to 1 at wavelengths shorter than about 4 μm and 0 at longer wavelengths.

Next we note that the flux of energy equal to one solar constant (1.37 kW m^{-2}) in this model falls entirely on the surface, raising its temperature in the absence of further energy input to about 250 K (section 1.3). The outgoing energy to space, here entirely from the atmosphere, must also be equal to the solar constant in order to achieve energy balance. However, the flux from the atmosphere occurs in both the upward and downward directions (figure 6), and so in equilibrium the surface receives a second contribution equal to that from the Sun, raising its temperature to $2^{0.25} \times 250$ or 297 K.

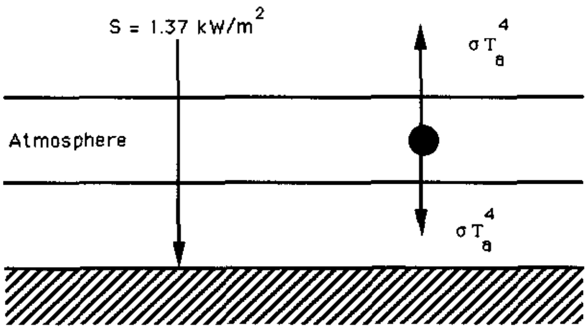

Figure 6. An elementary model of the greenhouse effect, in which the atmosphere is treated as a homogeneous layer of temperature T_a which is perfectly transparent to solar radiation and perfectly opaque in the thermal infrared. The surface receives the equivalent of two solar constants, raising its mean temperature from 250 to 297 K.

If the atmosphere were homogeneous in reality, this enhancement would be the upper limit on what could be achieved. In fact, it is clear (from studying Venus, for example) that the greenhouse enhancement can be not one but many solar constants. This becomes possible if we consider models in which the lower atmosphere is still opaque in the infrared, but warmer in its lower regions (which warm the surface) than its upper (which radiate to space).

3.1.2. A simple inhomogeneous model. A more sophisticated, but still basic, model can be formulated by introducing a more accurate representation of the vertical structure of the atmosphere, and by allowing for the spectral band structure in its opacity. It was explained above (section 2.1.2) how the lapse rate of the atmosphere in the troposphere and the radiative equilibrium temperature of the stratosphere can be obtained from elementary expressions obtained from first principles. These expressions (equations (1)–(8)) provide the basis for a simple model of the greenhouse effect which can be used to produce forecasts, albeit naive, of global change arising from an increased atmospheric CO_2 burden, and to demonstrate the role of key feedback processes, for example those involving clouds.

The model atmospheres in figure 7 are all characterised by a lapse rate Γ, a tropopause height z_t, and surface and stratospheric temperatures T_0 and T_x respectively. They differ only in the amount of CO_2 which is assumed to be present in the atmosphere, expressed as the mixing ratio by volume a_{CO_2} (ppmv), and in the value assumed for the mean albedo or reflectivity of the Earth. These models were computed in the following manner.

First make a number of simplifying assumptions:

1. The lapse rate of the troposphere and the temperature of the lower stratosphere are both constant. These are reasonable approximations to reality, and permit the 'climate' to be represented graphically as a temperature profile made up of various combinations of two straight lines with fixed gradients (Γ and zero, respectively).

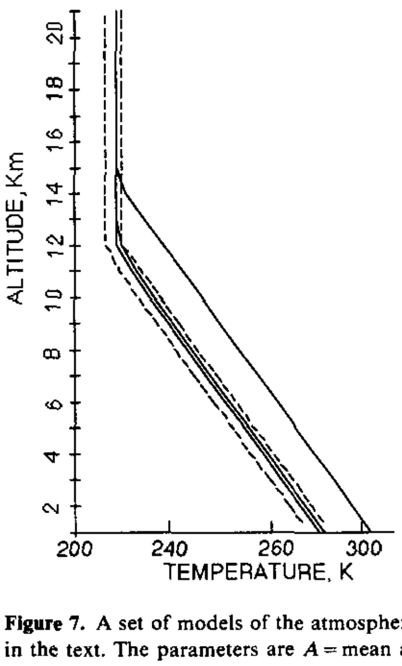

Figure 7. A set of models of the atmospheric temperature profile, calculated as described in the text. The parameters are A = mean albedo of the Earth, a_{CO_2} = abundance of CO_2 in ppmv. The full lines have A fixed at 0.3 and, from left to right, a_{CO_2} = 330, 350 and 660; the broken curves have a_{CO_2} fixed at 330 and, from left to right, A = 0.35 and 0.25.

2. The tropopause occurs at that level z_t where the optical depth, measured from the top of the atmosphere down, reaches unity. The physical reasoning assumed in adopting such a definition is that the atmosphere makes the transition from being optically thick, where vertical convection is the most efficient means of transferring

energy upwards, to becoming optically thin, where radiation takes over and the temperature becomes approximately constant, at the tropopause, for the reasons noted earlier.

3. The opacity of the atmosphere measured vertically is due principally to the strong absorption lines of species such as carbon dioxide and water vapour. For this kind of absorption, the optical depth is a function of the square root of the product of pressure p and absorber mass m (see, for example, Goody and Yung 1989). The path of absorber in the isothermal column above the tropopause at pressure p_t is Hap_t, where a is the mixing ratio by volume of the absorbing gas (assumed constant with height, a reasonable approximation for CO_2, which is overwhelmingly the most important infrared absorber in the stratosphere) and H is the atmospheric scale height (again constant with height, a fair approximation in the lower stratosphere). The optical depth, and hence the equivalent grey absorption coefficient, is therefore a function of the product of the pressure and the square root of the mixing ratio, so that:

$$\tau = f(p_t\sqrt{a}) = 1. \tag{13}$$

We do not need the form of the function f for present purposes, but see Goody and Yung (1989) for a discussion.

4. We assume the 'optically thin' approximation for the opacity of the stratosphere, which gives the value in (8) for its temperature, T_x.

5. The 'present day' atmospheric profile in figure 7 is based on the US Standard Atmosphere (1976), with the stratospheric temperature slightly reduced to be consistent with an albedo $a = 0.3$ according to (2) and (8). In the USSA, the tropopause occurs at a height of 11 km and a pressure of 227 mb, which are considered to be representative means of the present-day atmosphere. Table 3 lists these and other constants of the basic model.

Table 3. Constants used to calculate model temperature profiles, and the radiative equilibrium, stratospheric and surface temperatures which result for present-day CO_2 abundances.

Scale height	H	8 km
Distance, Earth to Sun	D_{ES}	1.50E+08 km
Radius of Sun	R_S	6.96E+05 km
Radius of Earth	R_E	6.38E+03 km
Effective temperature of Sun	T_S	5.79E+03 K
Solar constant	S	1.37E+03 W m^{-2}
Net sunfall	E	1.23E+11 W
Albedo	A	0.30
Effective temperature of Earth	T_E	255 K
Temperature of stratosphere	T_x	214.44 K
Temperature of surface	T_0	284.41 K

Now consider the effect of changing the absorber amount, including the case where the amounts are doubled, a standard case for climate modellers (the resulting change in T_0 is often called the *climate sensitivity*) and a situation which may occur on Earth (for CO_2 at least) during the next fifty or so years. Then, according to (13), p_t will move to $1/\sqrt{2}$ its present value, as indicated in table 4. This table also shows the effect of increasing CO_2 from its value around 1950 of 330 ppmv to its current level of about 350 ppmv. In both cases, the effect of increasing the absorber concentration is to raise

Table 4. Model temperature profiles for different CO_2 abundances, with A fixed at 0.3.

Approx. height (km)	Pressure p (mb)	Temp (K) for 330 ppmv	Temp (K) for 350 ppmv	Temp (K) for 660 ppmv
0	1000	284.41	285.94	302.44
1	882.49	277.91	279.44	295.94
2	778.80	271.41	272.94	289.44
3	687.28	264.91	266.44	282.94
4	606.53	258.41	259.94	276.44
5	535.26	251.91	253.44	269.94
6	472.36	245.41	246.94	263.44
7	416.86	238.91	240.44	256.94
8	367.87	232.41	233.94	250.44
9	324.65	225.91	227.44	243.94
10	286.50	219.41	220.94	237.44
11	252.84	214.44	215.97	230.94
12	223.13	214.44	214.44	224.44
13	196.91	214.44	214.44	217.94
14	173.77	214.44	214.44	214.44
15	153.35	214.44	214.44	214.44
16	135.33	214.44	214.44	214.44
17	119.43	214.44	214.44	214.44
18	105.39	214.44	214.44	214.44
19	93.01	214.44	214.44	214.44
20	82.08	214.44	214.44	214.44

the height of the tropopause to a lower pressure. Physically, what is happening is to increase the depth of the convective layer to offset the greater opacity of the atmosphere, so that heat is still brought up to the unit optical depth level where it can radiate to space (note that the model does not depend on the relationship $\tau = 1$ at the tropopause, any constant value of τ would have the same effect. However, it can be shown from simple radiative transfer theory that most of the heat reaches space from levels near $\tau = 1$). The stratospheric temperature remains unchanged because, in the optically thin approximation, this depends only on the equilibrium temperature of the planet, T_E.

The calculation for the smaller increase can be used to check how well this simple model performs, since we have at least a rough estimate of the global warming from compilations of surface temperature measurements made over the last 30 years (figure 8). There is a large amount of scatter in the data, due in part to the natural variability of the climate system, but also the difficulties inherent in forming such an average from disparate data. As we discuss elsewhere, this scatter is large enough that the data will not reveal any global warming with certainty until, and if, the effect becomes larger. For present purposes, however, let us attribute about 0.25 K of surface temperature increase to the increase of CO_2 from 330 to 350 ppmv, based on figure 8. This should be compared to the model increase of 1.53 K, about 6 times too large. The prediction of an 18 K increase as a result of doubling CO_2 is also too high by a factor of 6 to 9, relative to predictions from more sophisticated models (section 3.3). Part of the discrepancy can be attributed immediately to the fact that we have tacitly assumed that the grey absorption coefficient has changed in proportion to a change of CO_2 abundance, whereas in fact the mean absorption coefficient of the atmosphere is, as we have seen, due to contributions from many molecules, not all of which are increasing

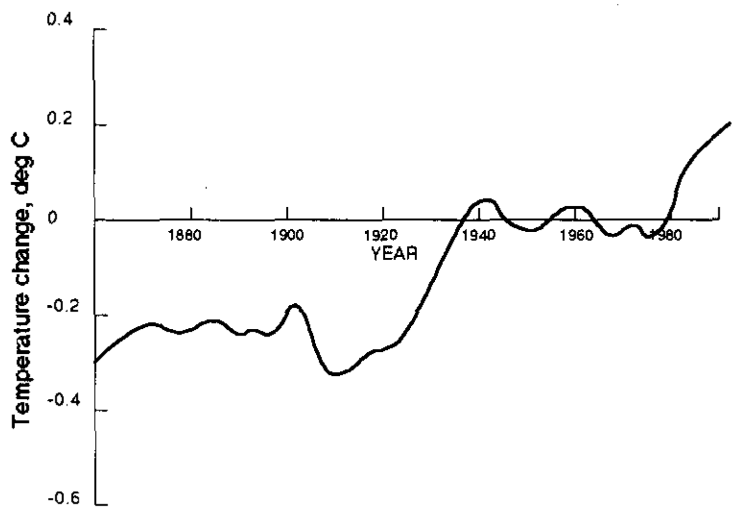

Figure 8. Global mean surface temperatures from 1861 to 1989, relative to the average for 1951 to 1980. From IPCC (1990).

at the same rate as CO_2 (although there is a fairly good correlation, see figure 2). IPCC (1990) estimates, again from sophisticated models, that about 55% of the warming in the last 10 years is due to CO_2. So, even if we now ignore changes in the non-CO_2 gases, our model estimate is still too high by a factor of 3.

Part of the discrepancy may be due to the difference between the 'committed' temperature increase (which we have calculated) and the 'realised' temperature change (which we measure). This difference is mainly due to the lag introduced by the thermal inertia of the oceans (section 2.5). IPCC (1990) estimates that the difference between committed and realised global mean temperature change is 50% if the climate sensitivity is 4.5 K and 20% if it is 1.5 K. This effect could therefore bring our model in line with the observations, certainly to within the error bars on the crude model and the noisy data.

3.2. Simple model with feedback

It will be clear from the discussions in section 2 that more accurate models need to include feedback processes. The two which have been identified as the most important are the increased humidity of the air which would follow an increase in mean surface temperature, and changes, probably increases, in the amounts of various types of clouds.

Although water vapour is the single most important absorber of those which produce the greenhouse effect, in our simple model, the effect of increasing water vapour is negligible provided the increase is confined to the troposphere. The physical reason is that the lower atmosphere is already optically thick, and the vertical transfer of energy is dominated not by radiation but by convection. This is represented by the fixed lapse rate Γ in our model. Stratospheric water vapour is another matter, but to first order this will not depend at all on the surface temperature because of the well-known cold-trapping effect on water of the low temperature at the tropopause. Carbon dioxide, on the other hand, is approximately uniformly mixed up to heights of around 60 km at least, so an increase in the mixing ratio raises the model tropopause height in the manner described above, but increasing tropospheric water makes no difference at all. (In reality, increased tropospheric water is important because the real atmosphere is not grey, and H_2O is an important absorber in the window regions. Still, like real-world modellers, we set this aside and press on.)

A second feedback process, this time probably tending to oppose any increase in the surface temperature, is the increase in the albedo of the Earth which results from an increase in cloudiness, itself postulated to result from increased humidity and convection in the troposphere. Clouds are generally more reflective than the surface they conceal, so increasing cloud cover tends to increase the albedo of the planet. Since albedo is one of the parameters used to formulate the simple model of the preceding section, this effect needs to be included, and permits us to examine the effects of changing albedo on the greenhouse warming predictions.

Let us assume that, for small changes, the albedo of the Earth depends linearly on the surface temperature, i.e. let $\mathrm{d}A/\mathrm{d}T_0 = k$ where k is a constant, assumed to be positive. The constant requires an erudite name to help disguise the naivety of the assumption it represents: we will call it the temperature-albedo feedback coefficient. With this parametrization included we find that the surface temperature increase of 1.5 K which results from a 20 ppmv increase in CO_2, reduces to a more reasonable value of 0.2 K, without invoking thermal inertia effects or partitioning between different greenhouse gases, if an albedo increase of 0.02 is invoked (table 5).

This corresponds to

$$k = \frac{\mathrm{d}A}{\mathrm{d}T_0} = \frac{0.02}{0.2} = 0.1\ K^{-1}. \tag{14}$$

Assuming rather boldly that the same linear approximation and the same value of the coefficient can be applied to the much larger CO_2-doubling scenario, and re-running

Table 5. Model temperature profiles for different albedos A, with CO_2 abundance fixed at 330 ppmv.

Approx. height (km)	Pressure p (mb)	Temperature (K) for $A = 0.25$	Temperature (K) for $A = 0.3$	Temperature (K) for $A = 0.35$
0	1000	288.15	284.41	280.48
1	882.49	281.65	277.91	273.98
2	778.80	275.15	271.41	267.48
3	687.28	268.65	264.91	260.98
4	606.53	262.15	258.41	254.48
5	535.26	255.65	251.91	247.98
6	472.36	249.15	245.41	241.48
7	416.86	242.65	238.91	234.98
8	367.87	236.15	232.41	228.48
9	324.65	229.65	225.91	221.98
10	286.50	223.15	219.41	215.48
11	252.84	216.65	214.44	210.51
12	223.13	216.65	214.44	210.51
13	196.91	216.65	214.44	210.51
14	173.77	216.65	214.44	210.51
15	153.35	216.65	214.44	210.51
16	135.33	216.65	214.44	210.51
17	119.43	216.65	214.44	210.51
18	105.39	216.65	214.44	210.51
19	93.01	216.65	214.44	210.51
20	82.08	216.65	214.44	210.51

the calculation, the cloud feedback effect reduces the predicted increase from 18 K to 3 K—again now in line with results from complex models—if the albedo increases to about 0.5. If a change of this size actually occurs, it may bring with it other changes, for example an increase in global mean precipitation.

A doubling of the Earth's cloud cover would be easy to detect; the smaller change from 0.3 to 0.32, which this model would require to have taken place over the last 20 years to match the observations, could easily pass undetected. It is not at all straightforward even to define what is meant by a mean cloud cover in the present sense, since albedo depends on quantities such as cloud height, particle size, geographical location and so forth. Recently, satellite experiments have been developed which measure the albedo and total thermal emission of the Earth from space, as will be discussed below (section 4.2). Even these could not detect changes in net albedo as small as 0.001 per year until several, perhaps as many as 10, years of data had been accumulated. Even then, the now-familiar problem of separating trends from slow fluctuations has to be surmounted.

In summary, then, even these very simple models give some insight into the principal processes at work, and allow us to go through the motions of accounting for recently observed small changes and of predicting major changes in the future. Obviously, the detailed values obtained from such a simple and in some ways arbitrary model are not very reliable, and some combination of the mechanisms described, plus some not considered, will in reality account for the difference between theory and observation in this case.

It is nevertheless reasonably certain that the basic physics implies non-trivial changes in mean surface temperature and/or global cloud cover in response to a doubling of the atmospheric CO_2 concentration. In the next section we will consider the extent to which this conclusion can be refined using sophisticated climate models.

3.3. Complex models

3.3.1. Formulation of complex models. Complex climate models are developed for two main reasons. The first is that by introducing more detailed physics, and more exact treatments of processes, we might hope eventually to achieve more reliable predictions over long periods of time. The second is to recognize that climate change is a regional affair and that changes in, for example, the mean surface temperature of the Earth will occur as a result of large increases in some regions with smaller, or even negative, increases in others. There will also be changes in the patterns of variability, not only of temperature but also of important quantities such as rainfall which meteorologists wish to understand and predict. Models with good spatial and temporal resolution, and comprehensive formulation or parametrization of radiative, dynamical and even chemical processes are required to address these goals.

Predictions obtained using existing climate models are only marginally plausible. Their estimates of the effect of doubling CO_2 have been reviewed by IPCC (1990). If the most reliable elements of the predictions are assumed to be those which are common to all the models, then the future will bring higher mean temperatures and increased precipitation, with the changes greater at high than at low latitudes. Southern Europe and central North America will warm up by more than the global average, and the monsoon season in Asia will intensify.

There are formidable obstacles to further improving these types of prediction. In spite of the rapid increase in the power of available computer hardware, it remains well short of being able to handle a fully detailed representation of the atmosphere by itself, let alone the complete climate system which must include the oceans, the cryosphere and perhaps the biosphere as well. This deficiency means that modellers rely on parametrizations of small-scale processes, often fairly crude and needing to be 'tuned' to reproduce the behaviour of the present atmosphere. More fundamentally, many atmospheric and oceanic processes (ozone photochemistry, for example) are not sufficiently well understood to allow their inclusion to sufficient accuracy even if adequate computer power were available. The net result is that no model can calculate from first principles a totally complete and accurate simulation of the present day climate, although the main features can be obtained fairly convincingly when tuning (in ways thought to be relatively insensitive to the boundary conditions, and therefore justifiable) is allowed.

Another problem with long-term predictions is that the necessary data inputs to the models—for example, the future rates at which man-made compounds will be added to the atmosphere—are all quite uncertain as well. Finally, we must raise the whole question of to what extent and over what scales of space and time a fundamentally chaotic system like the atmosphere is predictable at all.

None of these difficulties suggests that climate modelling is not worthwhile, but does indicate why it is that the results are always tagged as being highly uncertain, even when immensely complex and expensive models are brought to bear. Climate forecasting is a slow, long-term and gradual affair and deciding where progress can be made is as important a problem as any other. On the other hand, policy makers who have to control the global emissions of CO_2 and other greenhouse gases require and are entitled to the best predictions which scientists can provide—with error bars, of course. It is beyond the scope of this review to describe the workings of complex climate models in detail. All of them are based on general circulation models of one kind or another, increasingly atmospheric and oceanic GCMs coupled together. The basic building blocks of any GCM are the so-called primitive equations (e.g. Gill 1982), which express momentum balance, hydrostatic balance, conservation of mass and conservation of energy. To these must be added schemes for radiative transfer, cloud formation, surface interactions, and all the processes described in the preceding sections. These frequently involve the simplification, for example by linearization, of complex processes, and constantly demand improvement. The resulting models all give a heavily damped, essentially linear response to changing boundary conditions, which is fairly unrealistic especially for long-term forecasts, and explains why models cannot at present reproduce observed interannual variability.

The limitations of the most detailed climate models which exist at present are thought to be dominated by uncertainties in the feedback processes involving clouds and aerosols (sections 1.5, 2.4.3). Mitchell *et al* (1989) illustrated the scale of the cloud feedback problem by a model experiment in which the impact of doubling CO_2 was calculated for three different cloud model parameterizations. In the first, cloud amount was assumed to depend only on the atmospheric relative humidity, with no other change in the cloud characteristics. In this model, the cloud cover actually declined with increased CO_2, and a global warming of 5.2 K was predicted. In the second model, an attempt was made to include the physics of cloud formation explicitly, with a resulting warming of 2.7 to 3.2 K. Finally, a run with the radiative effects of water and ice particles included explicitly produced a warming of only 1.9 K.

3.3.2. Results from complex models. Here we draw on the results of two recent major studies which have used a diversity of models to try to bound the range of predictions admitted by the current state of the art in climate modelling. The Intergovernmental Panel on Climate Change (IPCC) of the World Meteorological Organisation (1990) adopted four 'scenarios' for the future emission of greenhouse gases. One of these, called 'business as usual', has the largest increases (a doubling of CO_2 by 2040, for example) while the others try to anticipate the effect of legislation to reduce emissions on the accumulation of greennhouse gases in the atmosphere. Figure 9(*a*) shows the range of predictions for mean global surface temperature for the 'business as usual' scenario, when the estimated uncertainties are taken into account. These uncertainties are largely attributed to the cloud feedback problem discussed above.

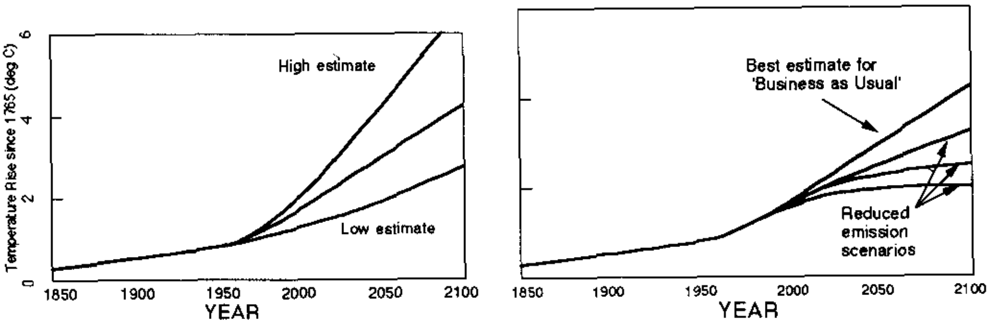

Figure 9. (*a*) Global mean temperature from 1850 to 2100, according to an evaluation by the Intergovernmental Panel on Climate Change of predictions from several of the best climate models (IPCC 1990). (*b*) Global mean temperature from 1850 to 2100, according to various scenarios which assume increasingly severe restrictions on the emission of man-made greenhouse gases. The 'best estimate' of the outcome of 'business as usual', from (*a*) is shown for comparison (IPCC 1990).

The slope of the most optimistic prediction for the 'business as usual' scenario is about 0.2 K per decade. The 'reduced emission' scenarios (figure 9(*b*)) suggeest that we are committed to a global warming of at least a degree during the next century, even with severe restraints on atmospheric pollution. The overall best estimate of the effect of doubling CO_2 is a 2.5 K global warming in the next 50 years.

The study commission of the Deutscher Bundestag (1989) reached similar conclusions, but with slightly higher values: a global temperature increase of 3 ± 1.5 K in the next few decades due to doubling of CO_2. They also introduce more extreme scenarios which admit global warmings of as much as 8.3 K or as little as 1.0 K by the year 2100. The former would correspond to a change in the global climate greater than that which the Earth experienced in the last great ice age 20 000 years ago. The latter is achievable, if the models are correct, only by imposing quite draconian limits on emissions, based on the proposition that a further 1 K of warming is the largest that civilization can readily tolerate.

Both surveys address the question of other changes, related to global warming. Sea level can be expected to rise, due to the thermal expansion of the oceans and, to a lesser extent, melting of glaciers and the polar icecaps. (The time constant for polar melting is so long that its effect is unlikely to be significant for 100 years or more). A rise of between 30 and 110 cm by 2100 AD is predicted by IPCC for the business-as-usual scenario, a mean rise of about 2 inches per decade. This could be halved by serious controls on emissions. Models indicate that extreme weather, deforestation and desertification could also be consequences of global warming although the uncertainty of specific predictions is very high.

4. Experimental studies

Progress with understanding the greenhouse effect, and planning to ameliorate the social and environmental effects of climate change, depends critically on our ability to design and carry out programmes of measurements. Theory and modelling alone cannot improve our understanding of the details of the physics on which models rely, and only good data can test and improve specific models and parameterizations. There are few direct experimental approaches to observing the greenhouse effect, however. We must probe the physical processes in the atmosphere and oceans at the same time as we monitor the climate system for variations and secular change.

In this section we consider a selection of the ways in which this is being done.

4.1. Ground-based programmes

Much of our existing body of knowledge about the climate system, and virtually all of the evidence, slender as yet, that global change may be beginning to occur, comes from more-or-less traditional instrumentation on ships, aircraft, and in meteorological stations around the world. With increasing interest in the global environment, efforts are being mounted to extend and coordinate such work; we consider three examples in the sections below.

It is worth prefacing the sections on *in situ* measurements by drawing attention to the fact that truly global quantities are very difficult to obtain from discrete data. For example, it might seem at first to be simple to obtain a mean global surface temperature by taking readings from meteorological stations distributed on a grid, and then forming a spatial and seasonal average. However, even today, this is very non-trivial, since the distribution of measurement is non-uniform in space and time and aliasing of seasonal and non-secular changes can easily occur. When long runs of data are required to search for trends, problems of changing measurement techniques and unreliable intercalibration are encountered. Measurements made within the urban 'heat island' are generally higher than those made at airports, etc, and the last fifty years has seen the gradual move of met stations from one to the other. And so on. The result is that, at the moment, those who believe that greenhouse warming has increased significantly since the industrial revolution are able to produce somewhat marginal but still quite convincing data (figure 7). Those more sceptical are also able to support their case with data (figure 10). It will probably be another decade at least before definite *experimental* evidence for or against global warming is in hand.

4.1.1. The network for the detection of stratospheric change (NSDC). The connection between trends in the stratosphere and global warming at the surface raises questions such as the following.

What is the actual rate of build-up of greenhouse gases in the atmosphere as a whole? Measurements at the surface are confused by local effects such as the concentration of pollutants near population centres, rainout and so forth.

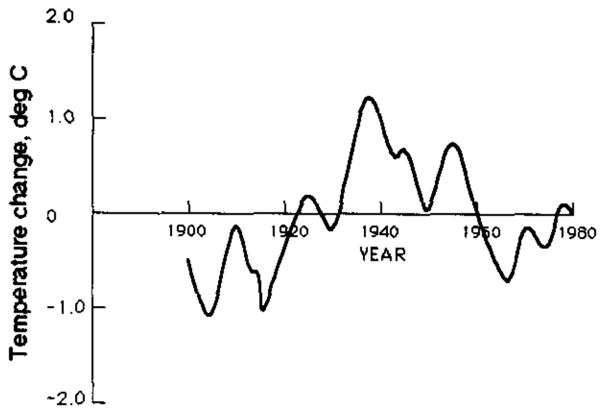

Figure 10. Annual mean temperature over the United States (after Karl *et al* 1988).

What role do chemical cycles play in determining the budget of trace gases? The stratosphere contains the main source for some greenhouse gases (e.g. ozone), and the main sink for others (e.g. methane). Predicting the rate of global change depends on understanding processes there.

Is global change occurring in the stratosphere? In the optically thin approximation (section 2.1.2), the mean stratospheric temperature depends only on the radiative equilibrium temperature on the planet, and is independent of changes in composition. In the real atmosphere, a few strong bands of CO_2 and O_3 are optically thick and this fact, plus thermodynamic effects, give rise to considerable temperature variability in the region. Most complex climate models predict some *cooling* of the stratosphere in response to a doubling of the CO_2 mixing ratio, indicating that enhanced thermal emission to other levels and to space, especially from the strong v_2 fundamental band near 15 μm, is expected to dominate over increased absorption of incoming solar or upwelling thermal radiation. On the other hand, measurements to date reveal fluctuations from year to year of around 10 K in zonally and monthly averaged stratospheric temperatures, but no measurable trend is discernable in the limited data obtained so far.

The NDSC consists of a series of stations being set up around the world. Each one has a similar set of remote-sensing instruments making regular observations of the stratosphere. At least six sites are planned, including one each in (or near) the Arctic and the Antarctic, plus sites at mid-latitude (the Alps), sub-tropical (Hawaii), southern hemisphere (New Zealand), and tropical locations. The instruments to be used, some of which are still under development, include the following.

Laser Radar Sounder (*lidar*) to obtain vertical profiles of temperature, aerosols and ozone. Ozone is measured using the differential absorption technique, which involves light of two different wavelengths, typically 0.335 and 0.308 μm, one of which is much more strongly absorbed by ozone than the other. The returned signals are very small but, given cloud-free skies, a three-hour integration time at night yields profiles from 18 to 440 km with vertical resolution of 200 m and absolute errors of less than 10%.

Temperature measurements by lidar rely on Rayleigh scattering from air molecules, typically at the wavelength of Nd lasers (0.532 μm). The signal strength is proportional to the air density and hence can be related to the temperature. The technique is valuable from about 30 to 70 km altitude, yielding temperature to an accuracy of better than

2 K with 200 m vertical resolution. Below 30 km, backscatter from atmospheric aerosols confuses the temperature signal and below about 26 km this contribution is so dominant that profiles of the aerosols themselves can be derived.

Infrared spectroscopy is the most versatile tool of the experimental atmospheric scientist. All of the key greenhouse gases, their source and sink species and their reaction partners have vibration-rotation bands in the 2–15 μm spectral region, all of which is accessible (in principle, simultaneously) to a Fourier-type spectrometer such as a Michelson interferometer. The instruments to be used by NDSC achieve spectral resolutions in the region of 0.002 cm^{-1}, which may provide limited vertical profile information and will separate the stratospheric from the (often much larger) tropospheric contribution to the spectral lines of interest. The Sun is used as the source, and is observed over a range of solar zenith angles. The range of air masses thus obtained makes it easier to observe weakly and more strongly absorbing species with the same set-up.

Ultraviolet and visible absorption spectroscopy. Short-wavelength spectrometers using the Sun or the Moon as a source can obtain column abundance measurements which complement or are superior to infrared data for certain species, including ozone, nitrogen dioxide, nitrogen trioxide, and chlorine and bromine oxides. This is an update of the classic Dobson technique, first developed for stratospheric ozone monitoring in the 1920s (Dobson 1930). Grating instruments with a spectral resolution on the order of 1 nm and integration times of 20 s to 20 min are used, and precisions of around 1% (for O_3 and NO_2) and around 10% for the other species, are claimed.

Microwave emission spectroscopy. Thermal emission from optically thin stratospheric thermal emission lines are monitored using microwave receivers operating at wavelengths around 22 GHz (for water vapour) to 279 GHz (for chlorine oxide). High spectral resolution defines the shapes of the pressure-broadened lines and allows vertical profiles to be retrieved, but with much less vertical resolution than is obtainable with active techniques such as lidar (around one scale height, 6–8 km). The most accessible species in the microwave are water (20–85 km), ozone (20–75 km), and chlorine oxide (25–45 km), with several others under consideration, including HCN, HO_2, CO, and N_2O.

4.1.2. Cloud studies using aircraft. The key problem involving clouds is to relate their formation processes, which can be parameterized in models, to their radiative effects. Much of the current work is being done using instrumented aircraft, which sample the cloud directly in order to measure its physical thickness, water content, temperature and pressure characteristics. At nearly the same time, radiometers on the aircraft can obtain the upward and downward fluxes of radiation above and below the cloud, in bandpasses corresponding to the solar and planetary fluxes (see also section 4.2.1).

Going beyond basic cloud characteristics to an understanding of the physical processes—principally scattering, emission and absorption—which occur in clouds, requires more sophisticated measurements. It is possible, for instance, to measure the size (and to some extent the shape if the particles are ice) of cloud droplets using optical methods whereby the drops are illuminated so that their shadows fall onto a detector array. Such data, if combined with multispectral measurements of the visible, near and far infrared flux in selected spectral intervals, allows the radiative properties of real clouds to be calculated and compared with the measured fluxes at its boundaries. Parametrizations useful in models can then be developed.

4.1.3. The biogeochemical ocean flux study (BOFS). BOFS is designed to elucidate the details of the oceanic carbon budget, in particular the exchange of CO_2 with the atmosphere (figure 4). Ship-borne measurements remain the only way to obtain data below the surface layer of the oceans, so BOFS addresses the above and other related questions by means of instrumented vessels sweeping across the North Atlantic. Particles are collected from many depths using towed, floating and moored sediment traps, pumps and coring drills. Water samples are analysed for their dissolved gas and material content. The data should establish how much of the CO_2 absorbed annually by the oceans is subsequently re-released, and how, where and when. It is known that some is retained permanently, after undergoing chemical and biological changes and sedimentation. A large amount is in solution, perhaps to be more freely released when global warming occurs (since the solubility of CO_2 in water decreases with increasing temperature) to provide an additional positive feedback mechanism. Minute plant life (phytoplankton) in the top 200 m of water consumes CO_2 through photosynthesis, but the variability of this process, its temperature dependence, and the ultimate fate of the CO_2 is obscure.

4.2. Satellite measurements

Observations from satellites are ideal for global atmospheric studies, since the whole planet can be surveyed regularly and efficiently, and in three dimensions. Most of the experimental effort now being brought to bear on an effort to better understand possible global change is in the form of large satellite programmes. Four of these, three originating in the USA and one in Europe, are described below.

4.2.1. Earth radiation budget experiment (ERBE).

Earth radiation budget experiment (ERBE) is a multi-satellite programme designed to measure the outgoing reflected solar radiation, i.e. the albedo, and the outgoing thermally emitted radiation, from Earth orbit. This is a less simple task than it may at first appear; the absolute calibration of accurate radiometers is notoriously difficult, particularly in space where the hostile environment can lead to deterioration of reference targets which cannot be checked. Furthermore, no real instrument can obtain a uniform spectral response over the wide wavelength ranges (roughly 0.4–4.0 μm for solar and 4.0–100 μm for thermal fluxes) required to measure local energy balance. Finally, albedo, although a simple concept, in practice is difficult to derive from data because integration over a 2π solid angle is required of a field which may have strong directional components (the thermal flux is easier because in this case it is reasonable to expect that cylindrical symmetry applies).

The instrumentation on the ERBE satellites addresses these difficulties by using redundant standard radiance targets, by breaking the wavelength range into segments which are measured and calibrated separately, and by using angular scans which later can be integrated into hemispherical fluxes. The last of these is problematical because the angular coverage over a given region from a single satellite obviously can never be complete; it is necessary to fit the data to empirical models of the reflectance properties of different types of surface and integrate the model, with a corresponding addition to the error budget.

The first ERBE satellite was launched in 1984. The results (Ramanathan *et al* 1989) provide at least tentative experimental evidence for the sign of the cloud feedback process: in the balance between increasing the opacity of the troposphere on one hand and increasing its albedo on the other, the latter dominates, resulting in net cooling.

With the obvious caveat that other important heat-transporting mechanisms were not being monitored at the same time, this supports the general theoretical arguments made above (sections 2.4.3, 3.3.1), although it is likely that the nature of the clouds themselves will change if the climate changes. It is possible that thicker, higher clouds could assist rather than reduce global warming, as they apparently do on Venus (section 4.3).

4.2.2. European remote sensing satellite (ERS-1). While the depths of the oceans are mostly inaccessible from space, the surface is easier to observe. The two most useful parameters for climate studies are the sea surface roughness, which is related to the surface wind speed and to momentum coupling between the atmosphere and the ocean, and the sea surface temperature, which reveals currents and controls the exchange of heat and moisture. The latest satellite system to attempt improved measurements of these parameters, ERS-1, uses a scatterometer for roughness measurements and a conical-scanning radiometer called ATSR (for along-track scanning radiometer). The main innovative feature of the latter is the use of near-simultaneous observations of the same spot on the ocean at two zenith angles $\theta=0$ (vertical) and $\theta=60°$. The measurements are made in the spectral window regions near 11 μm and 3.7 μm, where the residual absorption is mostly due to aerosols (and clouds when present), and continuum-type absorption by water vapour. Near-simultaneous measurements through two paths differing in air mass by a factor of 2 ($=\sec 60°$) allow the surface term to be separated from the atmospheric term in the radiative transfer equation for the radiance I_λ leaving the top of the atmosphere at a particular wavelength λ (Houghton *et al* 1985):

$$I_\lambda=\int_0^\infty B_\lambda(T)\frac{\mathrm{d}\tau_\lambda(\theta,z,\infty)}{\mathrm{d}z}\,\mathrm{d}z+\varepsilon_\lambda B_\lambda(T_0)\tau_\lambda(\theta,0,\infty). \tag{15}$$

In this expression B is the Planck blackbody function of temperature T at wavelength λ and $\tau_\lambda(\theta, z_1, z_2)$ is the transmission of the atmosphere at zenith angle θ between the two levels z_1 and z_2 at the same wavelength. T_0 is the surface temperature which is being sought, and ε_λ the surface emissivity, which is to a good approximation independent of θ for the ocean surface provided θ is significantly less than 90°. A term to allow for the reflectivity of the surface, normally small at these wavelengths, has been neglected in (15). Simulations suggest that ATSR can retrieve T_0 with an accuracy of 0.3 K, about 3 times better than at present.

The scientific significance of these accurate measurements is twofold: firstly, the flux of energy (and, by inference, moisture) into the atmosphere from the ocean can be obtained and changes monitored, and secondly ocean currents and associated phenomena with a recognised climate impact such as *El Nino* (Wells 1986) can be observed.

4.2.3. The upper atmosphere research satellite (UARS). UARS is the first of the new generation of very large Earth satellites dedicated to atmospheric measurements. Its objectives are to investigate and understand the complex interplay between radiation, dynamics and chemistry in the atmosphere, concentrating on altitudes ranging from 10 to 80 km (the middle, and not the upper, atmosphere in current nomenclature, the name of the mission notwithstanding). This is a popular region for atmospheric process studies, free of the complications of cloud, precipitation, and surface interactions, and so a 'flanking' manoeuvre in the general scientific attack on the greenhouse problem. UARS was conceived primarily to address the question of possible stratospheric ozone chemistry and possible ozone depletion, itself a relevant greenhouse issue.

The UARS spacecraft is approximately 11 m long and weighs nearly 7 tonnes. The scientific payload consists of nine large sensors: one microwave and three infrared spectrometers for atmospheric composition and temperature measurements, two Doppler wind-sending devices, and three instruments for measuring the solar spectral and particle fluxes into the atmosphere. The infrared sensors use different approaches with overlapping objectives; the combined data is expected to have the accuracy, coverage and verifiability needed to characterise temperature and composition fields and their fluctuations with latitude, season, solar activity and other parameters. The cryogenic limb array etalon spectrometer (CLAES) uses solid Fabry–Perot etalons, scanned in wavelength by tilting, to measure thermal emission spectra in selected parts of the mid-IR spectrum (4–16 μm). Very high sensitivity is obtained by cooling the entire spectrometer in a huge cryostat containing solid neon and solid CO_2. The improved stratospheric and mesospheric sounder (ISAMS) also measures thermal emission, but using pressure-modulated gas cells to select the spectral lines of CO_2, CO, H_2O, CH_4, and the oxides of nitrogen by the gas correlation technique (Houghton *et al* 1984). The halogen occultation experiment (HALOE) also uses gas correlation, but observes the atmosphere in transmission with the Sun as source. All three IR instruments, and the microwave limb sounder (MLS), observe the atmosphere at the limb to achieve high vertical resolution (around 2 km).

4.2.4. The Earth observing system (EOS). The successor to UARS is the Earth observation system or EOS, a giant polar-orbiting platform 16 m long, weighing over 13.6 tonnes. For present purposes, the most interesting feature of EOS is that it will extend the process studies of UARS down into the troposphere, where the interaction with the surface is greatest. We describe just two of the many instruments which are being considered for this purpose. Both face the various difficulties of measuring the atmosphere near the surface from space: fluctuations in the spectral properties of the surface, cloud contamination of radiances, and the fact that the limb-viewing technique cannot be used because the troposphere is too opaque when viewed tangentially. This last point means that the vertical resolution which can be obtained is poor, and the abundance retrievals are subject to errors introduced by the lack of knowledge of the shape of the vertical profile. Nevertheless, tropospheric studies clearly are an essential part of a long-term programme to understand climatic change and must be tackled.

MOPITT (for measurements of pollution in the troposphere) is a gas correlation instrument viewing vertically downwards and obtaining a small footprint a few km square on the surface. It uses the pressure modulated gas correlation technique, like ISAMS, to separate the spectral lines of its target species, in this case carbon monoxide and methane, from the forest of lines due to water vapour and other more abundant molecules. With high spatial resolution maps, the problem of finding cloud-free regions is reduced and MOPPITT offers the possibility of studying the production and subsequent evolution of concentrations of these two greenhouse gases.

TES (for tropospheric emission spectrometer) is an immensely sophisticated Fourier transform spectrometer with mechanically cooled optics and detectors, which has the potential to measure thermal emission from tropospheric minor constituents with high spectral and high spatial resolution, over a wide spectral range. Table 6 summarises the species which a design study (Beer *et al* 1990) suggests will be measurable. Global

Table 6. Illustrating the inventory of tropospheric species which is measurable by a state-of-the-art instrument on an Earth satellite—the tropospheric emission spectrometer planned for the Earth observing system around the turn of the century (after Beer *et al* 1990).

HO_x	NO_Y	Hydrocarbons	SO_x	Halocarbons (CFCs)	Others
H_2O	NO	CH_4	SO_2	$CFCl_2$	O_3
H_2O_2	NO_2	C_2H_6	COS	CF_2CL_2	CO
	HNO_3	C_2H_2			CO_2
	NH_3				N_2O

measurements of all of these species from space, from the surface up through the middle atmosphere, will surely advance the atmospheric chemistry part of the greenhouse puzzle rapidly once it is available. However, TES highlights another problem—the long development time for space measurement systems advanced enough to make the critical measurements needed to clarify the changes in the greenhouse effect. If NASA makes the decision to build TES now, it will not fly until early in the next century, by which time a further global warming of 0.2 to 0.5 K may have taken place.

4.3. Inferences from other planets

The greenhouse effect is not, of course, unique to the Earth. It plays a major role in determining the surface environment on the other terrestrial planets with atmospheres, Mars and Venus, and the same greenhouse gases (principally carbon dioxide and water vapour) are responsible. The existence of these planets offers an important opportunity to see how the greenhouse effect works in situations other than that we observe on the Earth.

Venus is a particularly interesting example, where an extreme case of the greenhouse effect raises the surface temperature to 730 K (higher than the melting point of lead). This happens in spite of the fact that the net solar input is significantly less than for the Earth, because the very high albedo of Venus (76% compared with around 30%) more than offsets its greater proximity to the Sun. This rather paradoxical situation occurs because Venus has a complete and deep cloud cover, and around a million times as much free carbon dioxide as the Earth. The fundamental reason for the former is probably the slow rotation rate of Venus, which improves the stability of the cloud layers. The latter arises probably through the absence of any surface water. As we have seen, much of the Earth's carbon dioxide is kept out of the atmosphere by dissolution and deposition in the oceans.

Nevertheless, the lesson of Venus is not simple to read. For instance, we can ponder the fact that the very reflective clouds allow a very hot surface, apparently in defiance of the principle incorporated in our simple terrestrial greenhouse model (section 3.1, and apparently confirmed by ERBE data, section 4.2.1) where clouds, by reflecting back the Sun's heat, cooled the whole troposphere. The reason in part for the difference is the fact that the Venusian clouds extend to great heights (around 60 km) above the surface, higher than the tropopause. The sulphuric acid droplets are also very opaque in the infrared. The cloud blanket therefore makes a large contribution to the backwarming effect, and this outweighs the albedo effect overall. It was at one time feared that the Earth could follow the same evolutionary track as Venus, with increasing temperatures, humidity and cloudiness following each other in a vicious spiral, but detailed model studies have now all but ruled that out.

Mars on the other hand has at least two rather different but no less striking statements to make. The first is the effect of airborne dust on temperatures. Figure 11 shows the results of comparisons between some early model calculations of the vertical temperature profile on Mars at different local times of day, and the actual range of measured profiles once these were actually obtained (by Mariner 9 in 1971, using the radio occultation technique). The difference turned out to be the increased infrared opacity of the atmosphere, caused by airborne dust. The great size of the effect—around 30 °C near the surface, and nearly 100 °C at the tropopause—emphasises the potential scale of the aerosol problem on the Earth, and the importance of dust-producing processes like volcanic eruptions and nuclear explosions, particularly if these occur on a global scale.

Mars also possesses the solar system's most dramatic evidence of past climate change. Today's frozen rocks and desert bear unmistakable features of rivers, lakes and seas. Mars must in the past have had a much thicker atmosphere, which now probably lies frozen beneath the surface and in the polar caps. Without this, it could not produce a greenhouse effect large enough to raise the temperature and pressure to values which could support liquid water on the surface. How could Mars have changed so much? The answer may lie in the high eccentricity of its orbit and the large fluctuations in sunfall which result from resonances between this and other orbital parammeters. This could in principle drive huge greenhouse fluctuations (Pollack *et al* 1987).

The climates of Mars and Venus clearly deserve much closer study. These close analogues of our own planet demonstrate extreme greenhouse behaviour while some terrestrial scientists are still prepared to argue that the climate of our own greenhouse-heated planet is essentially invariant. Recent space missions, most notably Pioneer Venus in 1978 and Mars Observer in 1992, have been designed to shed light on aspects of the climates of our planetary neighbours. Pioneer, in particular, by measurements of the albedo from orbit and of the physical properties of the atmosphere *in situ*, from descent probes, was able to show that the greenhouse effect does indeed account for

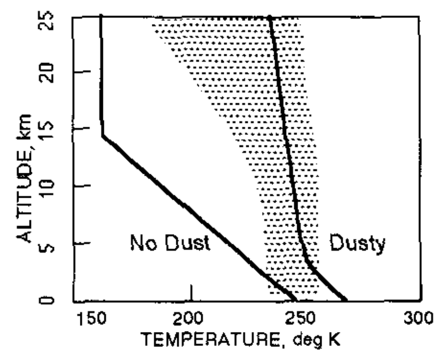

Figure 11. Temperature profiles for Mars showing the effect of direct solar heating due to dust absorption. The shaded areas show the range of Mariner 6 and 7 observations using the radio occultation technique. The curves on the left are calculated for a clear CO_2 atmosphere, while those on the right contain a representative amount of dust. (After Gierasch and Goody 1972).

the searing surface temperatures. Mars Observer will study the temperature, dynamics, dust and humidity variations, and energy balance of Mars' present day atmosphere, but a further dissection of its climate history is likely to have to await sophisticated robot, and eventually human, explorers in the next century.

5. Conclusion

There is no doubt at all that the greenhouse effect is a real and important component of the Earth's climate system. The way in which it works at the present time on Earth (and Venus, and Mars) is quite well understood in terms of radiative transfer theory and thermodynamics and is amenable to detailed calculation for a static case in which the basic atmospheric state (composition, cloud amount etc.) is well defined. These same calculations indicate that changes in the minor constituent inventory of the atmosphere will result in changes in the greenhouse warming of the surface, from its present value of about 33 K to some rather larger number. The current best estimate (IPCC 1990) is that the increase will fall in the range 0.2–0.5 °C per decade. More than a degree of accumulated increase is enough to have serious environmental and social effects.

Unfortunately, the calculations of which we are capable at present are much less reliable when it comes to predicting the effects of change than they are in accounting for the status quo. To some extent it is obvious why this should be so—most complex physical systems are like this. In the case of the climate system, comprising the atmosphere, oceans, cryosphere, biosphere, and the Sun, the coupling is so complex and so nonlinear that specific predictions are all but impossible. Nevertheless, it is possible to calculate the forcing, and this, with extrapolations from our best and most complete physics-based computer models, plus what we have learned about the experience of our planetary neighbours, suggests strongly that the response to current rates of pollution is likely to be significant change, in decades rather than centuries, nearly certainly in the direction of global warming.

The point at which the state of the art becomes really unreliable is in putting reasonably precise numbers on the rates at warming may occur, and regional variations really cannot be forecast at all, except maybe in terms of rough latitudinal dependences. There are also major difficulties, associated with the evolution in the amounts and detailed properties of clouds and aerosols and their role in a changing, and probably warmer and more polluted, world of the future. Scenarios can be constructed in which feedback mechanisms, probably involving cloud and aerosols, cancel out warming trends due to CO_2 and other substances in the gaseous phase, but no convincing scientific reasons have been published to explain why this particular balance should prevail over the predictions of our most complete physical models, which consistently predict warmings of 1.5 to 5 °C in response to a doubling of atmospheric CO_2 or its equivalent in combination with other greenhouse gases. On the other hand, climate modelling is a young enough discipline that the predictions are likely to evolve in the next decade or two, and drastic changes (for better or worse) in the expected trends still cannot be ruled out.

Better model parametrizations, larger, faster computers, and, perhaps above all, measurement programmes to study processes and test models, are being developed intensively at a number of institutes and slowly improving our confidence in the scale of the problem which mankind faces. Reasonably reliable predictions of at least the general trends the climate will follow in the next fifty years are probably not more than one or two decades away. In that time, it is also likely that the first irrefutable evidence for a current warming trend will be obtained, as measurements get better and the integrated effect gets larger. However, many aspects of global behaviour involving the atmosphere, such as long term weather forecasting, interannual variability or very long term trends, probably depend on a mixture of systematic changes and essentially chaotic behaviour. In these cases, we are still in the situation of needing to better understand which elements of climate are predictable and which are not.

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