

# CLIMATE CHANGE

*The IPCC Scientific Assessment*

WORLD METEOROLOGICAL ORGANIZATION / UNITED NATIONS ENVIRONMENT PROGRAMME

**INTERGOVERNMENTAL PANEL ON CLIMATE CHANGE**

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## Radiative Forcing of Climate

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# CONTENTS

<b>Executive Summary</b>	45	<b>2.3 Other Radiative Forcing Agents</b>	61
<b>2.1 Introduction</b>	47	2.3.1 Solar Radiation	61
<b>2.2 Greenhouse Gases</b>	47	2.3.1.1 Variability due to orbital changes	61
2.2.1 Introduction	47	2.3.1.2 Variability due to changes total in solar irradiance	61
2.2.2 Direct Effects	49	2.3.2 Direct Aerosol Effects	63
2.2.3 Indirect Effects	50	2.3.3 Indirect Aerosol Effects	64
2.2.4 Relationship Between Radiative Forcing and Concentration	51	2.3.4 Surface Characteristics	65
2.2.5 Past and Present Changes in Radiative Forcing	54	<b>2.4 The Relative Importance of Radiative Forcing Agents in the Future</b>	66
2.2.6 Calculations of Future Forcing	55		
2.2.7 A Global Warming Potential Concept for Trace Gases	58	<b>References</b>	67

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## EXECUTIVE SUMMARY

1 The climate of the Earth is affected by changes in radiative forcing due to several sources (known as radiative forcing agents) these include the concentrations of radiatively active (greenhouse) gases, solar radiation, aerosols and albedo. In addition to their direct radiative effect on climate, many gases produce indirect effects on global radiative forcing.

2 The major contributor to increases in radiative forcing due to increased concentrations of greenhouse gases since pre industrial times is carbon dioxide (CO<sub>2</sub>) (61%) with substantial contributions from methane (CH<sub>4</sub>) (17%), nitrous oxide (N<sub>2</sub>O) (4%) and chlorofluorocarbons (CFCs) (12%). Stratospheric water vapour increases, which are expected to result from methane emissions, contribute 6%, although evidence for changes in concentration is based entirely on model calculations.

The contribution from changes in tropospheric and stratospheric ozone is difficult to estimate, increased levels of tropospheric ozone may have caused 10% of the total forcing since pre industrial times. Decreases in lower stratospheric ozone may have decreased radiative forcing in recent decades.

3 The most recent decadal increase in radiative forcing is attributable to CO<sub>2</sub> (56%), CH<sub>4</sub> (11%), N<sub>2</sub>O (6%) and CFCs (24%), stratospheric H<sub>2</sub>O is estimated to have contributed 4%.

4 Using the scenario A ("business-as-usual" case) of future emissions derived by IPCC WG3, calculations show the following forcing from pre industrial values (and percentage contribution to total) by the year 2025:

CO<sub>2</sub> 2.9 Wm<sup>-2</sup> (63%), CH<sub>4</sub> 0.7 Wm<sup>-2</sup> (15%), N<sub>2</sub>O 0.2 Wm<sup>-2</sup> (4%), CFCs and HCFCs 0.5 Wm<sup>-2</sup> (11%), stratospheric H<sub>2</sub>O 0.2 Wm<sup>-2</sup> (5%)

The total, 4.6 Wm<sup>-2</sup> corresponds to an effective CO<sub>2</sub> amount of more than double the pre-industrial value.

5 An index is developed which allows the climate effects of the emissions of greenhouse gases to be compared. This is termed the Global Warming Potential (GWP). The GWP depends on the position and strength of the absorption bands of the gas, its lifetime in the atmosphere, its molecular weight and the time period over which the climate effects are of concern. A number of simplifications are used to derive values for GWPs and the values presented here should be considered as preliminary. It is quoted here as relative to CO<sub>2</sub>.

Over a 500 year time period, the GWP of equal mass emissions of the gases is as follows:

CO<sub>2</sub> 1, CH<sub>4</sub> 9, N<sub>2</sub>O 190, CFC-11 1500, CFC-12 4500, HCFC-22 510

Over a 20 year time period, the corresponding figures are:

CO<sub>2</sub> 1, CH<sub>4</sub> 63, N<sub>2</sub>O 270, CFC-11 4500, CFC-12 7100, HCFC-22 4100

Values for other gases are given in the text. There are many uncertainties associated with this analysis, for example the atmospheric lifetime of CO<sub>2</sub> is not well characterized. The GWPs can be applied by considering actual emissions of the greenhouse gases. For example, considering anthropogenic emissions of all gases in 1990, and integrating their effect over 100 years, shows that 60% of the greenhouse forcing from these emissions comes from CO<sub>2</sub>.

6 Although potential CFC replacements are less (or, in some cases, not at all) damaging to the ozone layer, the GWPs of several of them are still substantial, however, over periods greater than about 20 years most of the substitutes should have a markedly smaller impact on global warming than the CFCs they replace, assuming the same emissions.

7 Changes in climate forcing over the last century due to greenhouse gas increases are likely to have been much greater than that due to solar radiation. Although decadal variations of solar radiation can be comparable with greenhouse forcing, the solar forcing is not sustained and oscillates in sign. This limits the ability of the climate system to respond to the forcing. In contrast, the enhanced greenhouse effect causes a sustained forcing.

8 Stratospheric aerosols resulting from volcanic eruptions can cause a significant radiative forcing. A large eruption such as El Chichon can cause a radiative forcing, averaged over a decade about one third of (but the opposite sign to) the greenhouse gas forcing between 1980 and 1990. Regional and short term effects of volcanic eruptions can be even larger.

9 Man made sulphur emissions which have increased in the Northern Hemisphere over the last century affect radiative forcing by forming aerosols and influencing the radiative properties of clouds so as to cool the Earth. It is very difficult to

estimate the size of this effect, but it is conceivable that this radiative forcing has been of a comparable magnitude, but of opposite sign, to the greenhouse forcing earlier in this century; regional effects could even have been larger. The change in

forcing due to sulphur emissions in the future could be of either sign, as it is not known whether the emissions will increase or decrease.

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## 2.1 Introduction

The climate of the Earth has the potential to be changed on all timescales by the way in which shortwave radiation from the Sun is scattered and absorbed, and thermal infrared radiation is absorbed and emitted by the Earth-atmosphere system. If the climate system is in equilibrium, then the absorbed solar energy is exactly balanced by radiation emitted to space by the Earth and atmosphere. Any factor that is able to perturb this balance, and thus potentially alter the climate, is called a radiative forcing agent.

Of particular relevance to concerns about climate change are the changes in radiative forcing which arise from the increases in the concentration of radiatively active trace gases (greenhouse gases) in the troposphere and stratosphere described in Section 1. These changes in concentration will come about when their emissions or removal mechanisms are changed so that the atmospheric concentrations are no longer in equilibrium with the sources and sinks of the gas. The growing concentrations of greenhouse gases such as carbon dioxide, methane, chlorofluorocarbons and nitrous oxide are of particular concern. In addition, indirect effects on radiative forcing can result from molecules that may not themselves be greenhouse gases but which lead to chemical reactions which create greenhouse gases. For example, indirect effects are believed to be altering the distribution of stratospheric and tropospheric ozone.

Although water vapour is the single most important greenhouse gas, the effect of changes in its tropospheric concentration (which may arise as a natural consequence of the warming) is considered as a feedback to be treated in climate models, similarly changes in cloud amount or properties which result from climate changes will be considered as feedbacks. Both these factors are discussed in Section 3. Possible feedbacks between ocean temperature and dimethyl sulphide emissions, which may alter sulphate aerosol amounts, are also considered to be a feedback and will be considered in Section 3.

Other factors can alter the radiative balance of the planet. The most obvious of these is the amount of solar radiation reaching the Earth and this is known to vary on a wide range of time scales. The amount of solar radiation absorbed by the Earth-atmosphere system is determined by the extent to which the atmosphere and Earth's surface reflect the radiation (their albedo) and by the quantities of gases such as ozone and water vapour in the atmosphere. The albedo of the Earth's surface can be affected by changes in the land surface, e.g., desertification. The planetary albedo can be altered by changes in the amount of aerosol particles in the atmosphere, in the stratosphere the dominant source is from volcanic eruptions, while in the troposphere the source can be either natural or man-made.

The planetary albedo will also change if the properties of clouds are changed, for instance, if additional cloud condensation nuclei are provided by natural or man-made changes in aerosol concentrations. Changes in aerosol concentrations can also affect radiative forcing by their ability to absorb thermal infrared radiation.

Although all of the above factors will be considered in this section, the emphasis will be very strongly on the greenhouse gases, as they are likely to change radiative forcing over the next few decades by more than any other factor, natural or anthropogenic. They are also candidates for any policy action which may be required to limit global climate change. Obviously factors such as those related to emissions from volcanoes and the effects of solar variability are completely outside our control.

The purpose of this section is to use the information described in Section 1, on how the forcing agents themselves have changed in the past and how, based on a number of emission scenarios, they may change in the future. This information will then be used in climate models, later in the report, to show the climate and sea level consequences of the emission scenarios.

However, we can also use the estimates of radiative forcing from this section in their own right, by looking at the relative contribution from each of the agents - and in particular the greenhouse gases. The advantage of dealing with radiative forcing, rather than climate change itself, is that we can estimate the former with a great deal more certainty than we can estimate the latter. In the context of policy formulation, the relative importance of these agents is of major significance in assessing the effectiveness of response strategies. The radiative forcing is expressed as a change in flux of energy in  $\text{Wm}^{-2}$ .

In order to formulate policy on the possible limitations of greenhouse gas emissions (undertaken within IPCC by Working Group III), it is essential to know how abatement of the emissions of each of the trace gases will affect global climate forcing in the future. This information can then be used for calculations of the cost-effectiveness of reductions, e.g.  $\text{CO}_2$  emissions compared to  $\text{CH}_4$  emissions. There is no ideal index that can be used for each gas, but values of one index, the Global Warming Potential, are derived in this section. Research now under way will enable such indices to be refined.

## 2.2 Greenhouse Gases

### 2.2.1 Introduction

A typical global-average energy budget for the climate system shows that about half of the incident solar radiation (at wavelengths between 0.2 and 4.0  $\mu\text{m}$ ) is absorbed at the Earth's surface. This radiation warms the Earth's surface which then emits energy in the thermal infrared region (4-



100 $\mu$ m), constituents in the Earth's atmosphere are able to absorb this radiation and subsequently emit it both upwards to space and downwards to the surface. This downward emission of radiation serves to further warm the surface, this warming is known as the greenhouse effect.

The strength of the greenhouse effect can be gauged by the difference between the effective emitting temperature of the Earth as seen from space (about 255K) and the globally-averaged surface temperature (about 285K). The principal components of the greenhouse effect are the atmospheric gases (Section 2.2.2), clouds and aerosols also absorb and emit thermal infrared radiation but they also increase the planetary albedo, and it is believed that their net effect is to cool the surface (see Sections 3.3.4 and 2.3.2). Of the atmospheric gases, the dominant greenhouse gas is water vapour. If H<sub>2</sub>O was the only greenhouse gas present then the greenhouse effect of a clear sky mid-latitude atmosphere, as measured by the difference between the emitted thermal infrared flux at the surface and the top of the atmosphere, would be about 60-70% of the value with all gases included, by contrast, if CO<sub>2</sub> alone was present the corresponding value would be about 25% (but note that because of overlap between the absorption bands of different gases, such percentages are not strictly additive).

Here we are primarily concerned with the impacts of changing concentrations of greenhouse gases. A number of basic factors affect the ability of different greenhouse gases to force the climate system.

The absorption strength and the wavelength of this absorption in the thermal infrared are of fundamental importance in dictating whether a molecule can be an important greenhouse forcing agent, this effect is modified by both the existing quantities of that gas in the atmosphere and the overlap between the absorption bands and those of other gases present in the atmosphere.

The ability to build up significant quantities of the gas in the atmosphere is of obvious importance and this is dictated not only by the emissions of the gas, but also by its lifetime in the atmosphere. Further, these gases, as well as those that are not significant greenhouse gases can, via chemical reactions, result in products that are greenhouse gases.

In addition, the relative strength of greenhouse gases will depend on the period over which the effects of the gases are to be considered. For example, a short-lived gas which has a strong (on a kg-per-kg basis) greenhouse effect may, in the short term, be more effective at changing the radiative forcing than a weaker but longer-lived gas, over longer periods, however, the integrated effect of the weaker gas may be greater as a result of its persistence in the atmosphere.

From this introduction, it is clear that an assessment of the strength of greenhouse gases in influencing radiative

forcing depends on how that strength is measured. There are many possible approaches and it is important to distinguish between them.

Some of the more important indices that have been used as measures of the strength of the radiative forcing by greenhouse gases include

- i) **Relative molecular forcing.** This gives the relative forcing on a molecule-per-molecule basis of the different species. It is normally quoted relative to CO<sub>2</sub>. Since the forcing of some atmospheric species (most notably CO<sub>2</sub>, methane and nitrous oxide) is markedly non-linear in absorber amount, this relative forcing will be dependent on the concentration changes for which the calculations are performed. A small change in current atmospheric concentrations is generally used. This measure emphasises that the contributions of individual gases *must not* be judged on the basis of concentration alone. The relative molecular forcing will be considered in Section 2.2.4.
- ii) **Relative mass forcing.** This is similar to the relative molecular forcing but is relative on a kilogram per kilogram basis. It is related to the relative molecular forcing by the molecular weights of the gases concerned. It will also be considered in Section 2.2.4.
- iii) **Contribution of past, present and future changes in trace gas concentration.** This measure, which can either be relative or absolute, calculates the contribution to radiative forcing over some given period due to observed past or present changes, or scenarios of future changes in trace gas concentration. This is an important baseline. The relative measures (i) and (ii) above, can belittle the influence of carbon dioxide since it is relatively weak on a molecule-per-molecule basis, or a kg-per-kg basis. This measure accounts for the fact that the concentration changes for CO<sub>2</sub> are between two and four orders of magnitude greater than the changes of other important greenhouse gases. This measure will be considered in sections 2.2.5 and 2.2.6. Care must be taken in interpreting this measure as it is sometimes presented as the total change in forcing since pre-industrial times and sometimes as the change in forcing over a shorter period such as a decade or 50 years.
- iv) **Global Warming Potential (GWP).** All the above measures are based on *concentration* changes in the atmosphere, as opposed to *emissions*. Assessing the potential impact of future emissions may be far more important from a policy point of view. Such measures combine calculations of the absorption strength of a molecule with assessments of its atmospheric lifetime, it can also include the indirect greenhouse effects due to chemical changes in the

atmosphere caused by the gas. The development of an index is still at an early stage, but progress has been made and preliminary values are given in Section 2.2.7.

A detailed assessment of the climatic effects of trace gases was made by WMO (1985) (see also Ramanathan et al., 1987). The effect of halocarbons has been considered in detail in the recent Scientific Assessment of Stratospheric Ozone (UNEP, 1989) (see also Fisher et al., 1990). This section should be considered as building on these assessments and bringing them up to date.

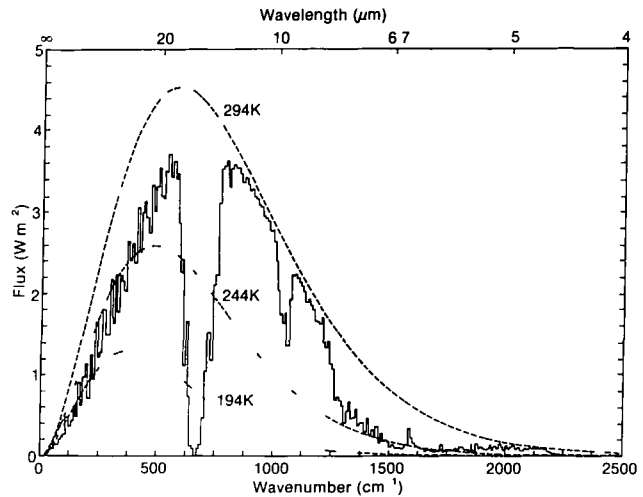
### 2.2.2 Direct Effects

Many molecules in the atmosphere possess pure-rotation or vibration-rotation spectra that allow them to emit and absorb thermal infrared radiation (4–100  $\mu\text{m}$ ), such gases include water vapour, carbon dioxide and ozone (but not the main constituents of the atmosphere, oxygen or nitrogen). These absorption properties are directly responsible for the greenhouse effect.

It is not the change in thermal infrared flux at the surface that determines the strength of the greenhouse warming. The surface, planetary boundary layer and the free troposphere are tightly coupled via air motions on a wide range of scales so that in a global-mean sense they must be considered as a single thermodynamic system. As a result it is the change in the radiative flux at the *tropopause*, and not the surface, that expresses the radiative forcing of climate system (see e.g., Ramanathan et al., 1987).

A number of factors determine the ability of an added molecule to affect radiative forcing and in particular the spectral absorption of the molecule in relation to the spectral distribution of radiation emitted by a black-body. The distribution of emitted radiation with wavelength is shown by the dashed curves for a range of atmospheric temperatures in Figure 2.1. Unless a molecule possesses strong absorption bands in the wavelength region of significant emission, it can have little effect on the net radiation.

These considerations are complicated by the effect of naturally occurring gases on the spectrum of net radiation at the tropopause. Figure 2.1 shows the spectral variation of the net flux at the tropopause for a clear-sky mid-latitude profile. For example, the natural quantities of carbon dioxide are so large that the atmosphere is very opaque over short distances at the centre of its 15  $\mu\text{m}$  band. At this wavelength the radiation reaching the tropopause, from both above and below, comes from regions at temperatures little different to the tropopause itself. The net flux is thus close to zero. The addition of a small amount of gas capable of absorbing at this wavelength has negligible effect on the net flux at the tropopause. The effect of added carbon dioxide molecules is, however, significant at the



**Figure 2.1:** The dashed lines show the emission from a black body ( $\text{W m}^{-2}$  per  $10 \text{ cm}^{-1}$  spectral interval) across the thermal infrared for temperatures of 294K, 244K and 194K. The solid line shows the net flux at the tropopause ( $\text{W m}^{-2}$ ) in each  $10 \text{ cm}^{-1}$  interval using a standard narrow band radiation scheme and a clear-sky mid-latitude summer atmosphere with a surface temperature of 294K (Shine, pers. comm.). In general, the closer this line is to the dashed line for 294K, the more transparent the atmosphere.

edges of the 15  $\mu\text{m}$  band, and in particular around 13.7 and 16  $\mu\text{m}$ . At the other extreme, in more transparent regions of the atmosphere (for example between 10 and 12  $\mu\text{m}$ ), much of the radiation reaching the tropopause from beneath is, for clear skies, from the warm surface and the lower troposphere; this emission is not balanced by downward emission of radiation from the overlying stratosphere. A molecule able to absorb in such a transparent spectral region is able to have a far larger effect.

The existing concentrations of a particular gas dictate the effect that additional molecules of that gas can have. For gases such as the halocarbons, where the naturally occurring concentrations are zero or very small, their forcing is close to linear in concentration for present-day concentrations. Gases such as methane and nitrous oxide are present in such quantities that significant absorption is already occurring and it is found that their forcing is approximately proportional to the square root of their concentration. Furthermore, there is significant overlap between some of the infrared absorption bands of methane and nitrous oxide which must be carefully considered in calculations of forcing. For carbon dioxide, as has already been mentioned, parts of the spectrum are already so opaque that additional molecules of carbon dioxide are even less effective; the forcing is found to be logarithmic in concentration. These effects are reflected in the empirical expressions used to calculate the radiative forcing that are discussed in Section 2.2.4.

A further consideration is the spectroscopic strength of the bands of molecules which dictates the strength of the infra-red absorption. Molecules such as the halocarbons have bands with intensities about an order of magnitude or more greater, on a molecule-per-molecule basis, than the 15  $\mu\text{m}$  band of carbon dioxide. The actual absorptance by a band is, however, a complicated function of both absorber amount and spectroscopic strength so that these factors cannot be considered entirely in isolation.

### 2.2.3 Indirect Effects

In addition to their direct radiative effects, many of the greenhouse gases also have indirect radiative effects on climate through their interactions with atmospheric chemical processes. Several of these interactions are shown in Table 2.1

For example, both atmospheric measurements and theoretical models indicate that the global distribution of ozone in the troposphere and stratosphere is changing as a result of such interactions (UNEP, 1989, also see Section 1)

Ozone plays an important dual role in affecting climate. While  $\text{CO}_2$  and other greenhouse gases are relatively well-

mixed in the atmosphere, the climatic effect of ozone depends strongly on its vertical distribution throughout the troposphere and stratosphere, as well as on its total amount in the atmosphere. Ozone is a primary absorber of solar radiation in the stratosphere where it is directly responsible for the increase in temperature with altitude. Ozone is also an important absorber of infrared radiation. It is the balance between these radiative processes that determines the net effect of ozone on climate. Changes in ozone in the upper troposphere and lower stratosphere (below 25 km) are most effective in determining the change in radiative forcing, with increased ozone leading to an increased radiative forcing which would be expected to warm the surface (e.g., Wang and Sze, 1980, Lacis et al., 1990). This is because the greenhouse effect is directly proportional to the temperature contrast between the level of emission and the levels at which radiation is absorbed. This contrast is greatest near the tropopause where temperatures are at a minimum compared to the surface. Above about 30 km, added ozone causes a decrease in surface temperature because it absorbs extra solar radiation, effectively robbing the troposphere of direct solar energy that would otherwise warm the surface (Lacis et al., 1990).

**Table 2.1:** Direct radiative effects and indirect trace gas chemical-climate interactions (based on Wuebbles et al., 1989)

Gas	Greenhouse Gas	Is its tropospheric concentration affected by chemistry?	Effects on tropospheric chemistry? *	Effects on * stratospheric chemistry?
$\text{CO}_2$	Yes	No	No	Yes, affects $\text{O}_3$ (see text)
$\text{CH}_4$	Yes	Yes, reacts with OH	Yes, affects OH, $\text{O}_3$ and $\text{CO}_2$	Yes, affects $\text{O}_3$ and $\text{H}_2\text{O}$
CO	Yes, but weak	Yes, reacts with OH	Yes, affects OH, $\text{O}_3$ and $\text{CO}_2$	Not significantly
$\text{N}_2\text{O}$	Yes	No	No	Yes, affects $\text{O}_3$
$\text{NO}_x$	Yes	Yes, reacts with OH	Yes, affects OH and $\text{O}_3$	Yes, affects $\text{O}_3$
CFC-11	Yes	No	No	Yes, affects $\text{O}_3$
CFC-12	Yes	No	No	Yes, affects $\text{O}_3$
CFC-113	Yes	No	No	Yes, affects $\text{O}_3$
HCFC-22	Yes	Yes, reacts with OH	No	Yes, affects $\text{O}_3$
$\text{CH}_3\text{CCl}_3$	Yes	Yes, reacts with OH	No	Yes, affects $\text{O}_3$
$\text{CF}_2\text{ClBr}$	Yes	Yes, photolysis	No	Yes, affects $\text{O}_3$
$\text{CF}_3\text{Br}$	Yes	No	No	Yes, affects $\text{O}_3$
$\text{SO}_2$	Yes, but weak	Yes, reacts with OH	Yes, increases aerosols	Yes, increases aerosols
$\text{CH}_3\text{SCH}_3$	Yes, but weak	Yes, reacts with OH	Source of $\text{SO}_2$	Not significantly
$\text{CS}_2$	Yes, but weak	Yes, reacts with OH	Source of COS	Yes, increases aerosols
COS	Yes, but weak	Yes, reacts with OH	Not significant	Yes, increases aerosols
$\text{O}_3$	Yes	Yes	Yes	Yes

\* - Effects on atmospheric chemistry are limited to effects on constituents having a significant influence on climate

Stratospheric water vapour is an important greenhouse gas. A major source of stratospheric water vapour is the oxidation of methane (e.g., WMO 1985), it is anticipated that increased atmospheric concentrations of methane will lead to increases in stratospheric water vapour. It is also possible that changes in climate will affect the transfer of water vapour from the troposphere to the stratosphere, although the sign of the net effect on stratospheric water vapour is unclear. Unfortunately, observations of stratospheric water vapour are inadequate for trend detection. In this section the impact of increased emissions of methane on stratospheric water vapour will be included as an indirect radiative forcing due to methane.

The oxidation of fossil based methane and carbon monoxide in the atmosphere lead to the production of additional carbon dioxide. Although CO<sub>2</sub> has no known chemical interactions of consequence within the troposphere or stratosphere its increasing concentrations can affect the concentrations of stratospheric ozone through its radiative cooling of the stratosphere. In the upper stratosphere the cooling slows down catalytic ozone destruction and results in a net increase in ozone, where heterogeneous ozone destruction is important, as in the Antarctic lower stratosphere ozone destruction may be accelerated by this cooling (UNEP, 1989). The combination of these indirect effects, along with their direct radiative effects, determines the actual changes in radiative forcing resulting from these greenhouse gases.

The hydroxyl radical, OH, is not itself a greenhouse gas but it is extremely important in the troposphere as a chemical scavenger. Reactions with OH largely control the atmospheric lifetime, and, therefore the concentrations of many gases important in determining climate change. These gases include CH<sub>4</sub>, CO, the non-methane hydrocarbons (NMHCs), the hydrochlorofluorocarbons (HCFCs), the hydrofluorocarbons (HFCs), CH<sub>3</sub>CCl<sub>3</sub>, H<sub>2</sub>S, SO<sub>2</sub> and dimethyl sulphide (DMS). Their reaction with OH also affects the production of tropospheric ozone, as well as determining the amounts of these compounds reaching the stratosphere, where these species can cause changes in the ozone distribution. In turn the reactions of these gases with OH also affects its atmospheric concentration. The increase in tropospheric water vapour concentration expected as a result of global warming would also increase photochemical production of OH. It is important that effects of interaction between OH and the greenhouse gases, along with the resulting impact on atmospheric lifetimes of these gases, be accounted for in analysing the possible state of future climate.

The indirect effects can have a significant effect on the total forcing. These effects will be detailed later in the section.

### 2.2.4 Relationship Between Radiative Forcing and Concentration

To estimate climate change using simple energy balance climate models (see Section 6) and in order to estimate the relative importance of different greenhouse gases in past, present and future atmospheres (e.g., using Global Warming Potentials, see Section 2.2.7), it is necessary to express the radiative forcing for each particular gas in terms of its concentration change. This can be done in terms of the changes in net radiative flux at the tropopause

$$\Delta F = f(C_0, C)$$

where  $\Delta F$  is the change in net flux (in  $\text{Wm}^{-2}$ ) corresponding to a volumetric concentration change from  $C_0$  to  $C$ .

Direct-effect  $\Delta F$ - $\Delta C$  relationships are calculated using detailed radiative transfer models. Such calculations simulate the complex variations of absorption and emission with wavelength for the gases included, and account for the overlap between absorption bands of the gases, the effects of clouds on the transfer of radiation are also accounted for.

As was discussed in Section 2.2.2, the forcing is given by the change in net flux at the tropopause. However as is explained by Ramanathan et al (1987) and Hansen et al (1981) great care must be taken in the evaluation of this change. When absorber amount varies, not only does the flux at the tropopause respond, but also the overlying stratosphere is no longer in radiative equilibrium. For some gases, and in particular CO<sub>2</sub>, the concentration change acts to cool the stratosphere, for others, and in particular the CFCs, the stratosphere warms (see e.g. Table 5 of Wang et al (1990)). Calculations of the change in forcing at the tropopause should allow the stratosphere to come into a new equilibrium with this altered flux divergence, while tropospheric temperatures are held constant. The consequent change in stratospheric temperature alters the downward emission at the tropopause and hence the forcing. The  $\Delta F$ - $\Delta C$  relationships used here implicitly account for the stratospheric response. If this point is ignored, then the same change in flux at the tropopause from different forcing agents can lead to a different tropospheric temperature response. Allowing for the stratospheric adjustment means that the temperature response for the same flux change from different causes are in far closer agreement (Lacis, personal communication).

The form of the  $\Delta F$ - $\Delta C$  relationship depends primarily on the gas concentration. For low/moderate/high concentrations, the form is well approximated by a linear/square-root/logarithmic dependence of  $\Delta F$  on concentration. For ozone, the form follows none of these because of marked vertical variations in absorption and concentration. Vertical variations in concentration change

**Table 2.2:** Expressions used to derive radiative forcing for past trends and future scenarios of greenhouse gas concentrations

TRACE GAS	RADIATIVE FORCING APPROXIMATION GIVING $\Delta F$ IN $Wm^{-2}$	COMMENTS
Carbon dioxide	$\Delta F = 6.3 \ln (C/C_0)$ where C is CO <sub>2</sub> in ppmv for C < 1000 ppmv	Functional form from Wigley (1987), coefficient derived from Hansen et al (1988)
Methane	$\Delta F = 0.036 (\sqrt{M} - \sqrt{M_0}) - (f(M, N_0) - f(M_0, N_0))$ where M is CH <sub>4</sub> in ppbv and N is N <sub>2</sub> O in ppbv Valid for M < 5ppmv	Functional form from Wigley (1987), coefficient derived from Hansen et al (1988) Overlap term, f(M, N) from Hansen et al (1988)*
Nitrous Oxide	$\Delta F = 0.14 (\sqrt{N} - \sqrt{N_0}) - (f(M_0, N) - f(M_0, N_0))$ with M and N as above Valid for N < 5ppmv	Functional form from Wigley (1987), coefficient derived from Hansen et al (1988) Overlap term from Hansen et al (1988)*
CFC-11	$\Delta F = 0.22 (X - X_0)$ where X is CFC-11 in ppbv Valid for X < 2ppbv	Based on Hansen et al (1988)
CFC-12	$\Delta F = 0.28 (Y - Y_0)$ where Y is CFC-12 in ppbv Valid for Y < 2ppbv	Based on Hansen et al (1988)
Stratospheric water vapour	$\Delta F = 0.011 (\sqrt{M} - \sqrt{M_0})$ where M is CH <sub>4</sub> in ppbv	Stratospheric water vapour forcing taken to be 0.3 of methane forcing without overlap based on Wuebbles et al (1989)
Tropospheric ozone	$\Delta F = 0.02 (O - O_0)$ where O is ozone in ppbv	Very tentative illustrative parameterization based on value from Hansen et al (1988)
Other CFCs, HCFCs and HFCs	$\Delta F = A (Z - Z_0)$ where A based on forcing relative to CFC-11 in Table 2.4 and Z is constituent in ppbv	Coefficients A derived from Fisher et al (1990)

\* Methane-Nitrous Oxide overlap term

$f(M, N) = 0.47 \ln [1 + 2.01 \times 10^{-5} (MN)^{0.75} + 5.31 \times 10^{-15} M (MN)^{1.52}]$ , M and N are in ppbv

Note typographical error on page 9360 of Hansen et al (1988) 0.014 should be 0.14

for ozone make it even more difficult to relate  $\Delta F$  to concentration in a simple way

The actual relationships between forcing and concentration derived from detailed models can be used to develop simple expressions (e.g., Wigley, 1987, Hansen et al 1988) which are then more easily used for a large number of calculations. Such simple expressions are used in this Section. The values adopted and their sources are given in Table 2.2. Values derived from Hansen et al have

been multiplied by 3.35 (Lacis, personal communication) to convert forcing as a temperature change to forcing as a change in net flux at the tropopause after allowing for stratospheric temperature change. These expressions should be considered as global mean forcings, they implicitly include the radiative effects of global mean cloud cover.

Significant spatial variations in  $\Delta F$  will exist because its value for any given  $\Delta C$  depends on the assumed

temperature and water vapour profiles. Variations will also occur due to spatial variations in mean cloudiness. These factors can produce marked differences in the relative contributions of different greenhouse gases to total radiative forcing in different regions but these are not accounted for here.

Uncertainties in  $\Delta F/\Delta C$  relationships arise in three ways. First, there are still uncertainties in the basic spectroscopic data for many gases. In particular, data for CFCs, HFCs and HCFCs are probably only accurate to within  $\pm 10\text{-}20\%$ . Part of this uncertainty is related to the temperature dependence of the intensities, which is generally not known. For some of these gases, only cross-section data are available. For the line intensity data that do exist, there have been no detailed intercomparisons of results from

different laboratories. Further information on the available spectroscopic data is given by Husson (1990).

Second, uncertainties arise through details in the radiative transfer modelling. Intercomparisons made under the auspices of WCRP (Luther and Fouquart, 1984) suggest that these uncertainties are around  $\pm 10\%$  (although schemes used in climate models disagreed with detailed calculations by up to 25% for the flux change at the tropopause on doubling  $\text{CO}_2$ ).

Third, uncertainties arise through assumptions made in the radiative model with regard to the following:

- (i) the assumed or computed vertical profile of the concentration change. For example, for CFCs and HCFCs, results can depend noticeably on the assumed change in stratospheric concentration (see e.g., Ramanathan et al., 1985).
- (ii) the assumed or computed vertical profiles of temperature and moisture.

**Table 2.3:** Radiative forcing relative to  $\text{CO}_2$  per unit molecule change, and per unit mass change in the atmosphere for present day concentrations.  $\text{CO}_2$ ,  $\text{CH}_4$  and  $\text{N}_2\text{O}$  forcings from 1990 concentrations in Table 2.5

TRACE GAS	$\Delta F/\Delta C$ per molecule relative to $\text{CO}_2$	$\Delta F/\Delta C$ per unit mass relative to $\text{CO}_2$
$\text{CO}_2$	1	1
$\text{CH}_4$	21	58
$\text{N}_2\text{O}$	206	206
CFC-11	12400	3970
CFC-12	15800	5750
CFC-113	15800	3710
CFC-114	18300	4710
CFC-115	14500	4130
HCFC-22	10700	5440
$\text{CCl}_4$	5720	1640
$\text{CH}_3\text{CCl}_3$	2730	900
$\text{CF}_3\text{Br}$	16000	4730
Possible CFC substitutes		
HCFC-123	9940	2860
HCFC-124	10800	3480
HFC-125	13400	4920
HFC-134a	9570	4130
HCFC-141b	7710	2900
HCFC-142b	10200	4470
HFC-143a	7830	4100
HFC-152a	6590	4390

**Table 2.4:** Radiative forcing of a number of CFCs, possible CFC substitutes and other halocarbons relative to CFC-11 per unit molecule and per unit mass change. All values, except  $\text{CF}_3\text{Br}$ , from Fisher et al., 1990.  $\text{CF}_3\text{Br}$  from Ramanathan et al., 1985.

TRACE GAS	$\Delta F/\Delta C$ per molecule relative to CFC11	$\Delta F/\Delta C$ per unit mass relative to CFC11
CFC-11	1.00	1.00
CFC-12	1.27	1.45
CFC-113	1.27	0.93
CFC-114	1.47	1.18
CFC-115	1.17	1.04
HCFC-22	0.86	1.36
HCFC-123	0.80	0.72
HCFC-124	0.87	0.88
HFC-125	1.08	1.24
HFC-134a	0.77	1.04
HCFC-141b	0.62	0.73
HCFC-142b	0.82	1.12
HFC-143a	0.63	1.03
HFC-152a	0.53	1.10
$\text{CCl}_4$	0.46	0.45
$\text{CH}_3\text{CCl}_3$	0.22	0.23
$\text{CF}_3\text{Br}$	1.29	1.19

- (iii) assumptions made with regard to cloudiness. Clear sky  $\Delta F$  values are in general 20% greater than those using realistic cloudiness
- (iv) the assumed concentrations of other gases (usually, present-day values are used). These are important because they determine the overall IR flux and because of overlap between the absorption lines of different gases
- (v) the indirect effects on the radiative forcing due to chemical interactions as discussed in Section 2.2.3

The overall effect of this third group of uncertainties on  $\Delta F$  is probably at least  $\pm 10\%$

Direct radiative forcing changes for the different greenhouse gases can be easily compared using the above  $\Delta F$ - $\Delta C$  relationships. There are two ways in which these comparisons may be made, per unit volumetric concentration change (equivalent to per molecule) or per unit mass change. Comparison for the major greenhouse gases are given in Table 2.3. The relative strength of the CFCs, HFCs and HCFCs, relative to CFC-11, are shown in Table 2.4 (from Fisher et al., 1990). It can be seen that, by

these measures, many of the potential CFC substitutes are strong infrared absorbers

### 2.2.5 Past and Present Changes in Radiative Forcing

Based on the expressions given in Table 2.2 the radiative forcing between 1765 and 1990 was calculated using observed variations of the greenhouse gases. The concentrations are given in Table 2.5, they are updated values from Wigley (1987) and Section 1. Values for 1990 have been extrapolated from recent values. In addition to the well-observed variations in the gases given in Table 2.5, it is assumed that increased concentrations of methane have led to increases in stratospheric water vapour, although such changes are based entirely on model estimates (see Section 2.2.3).

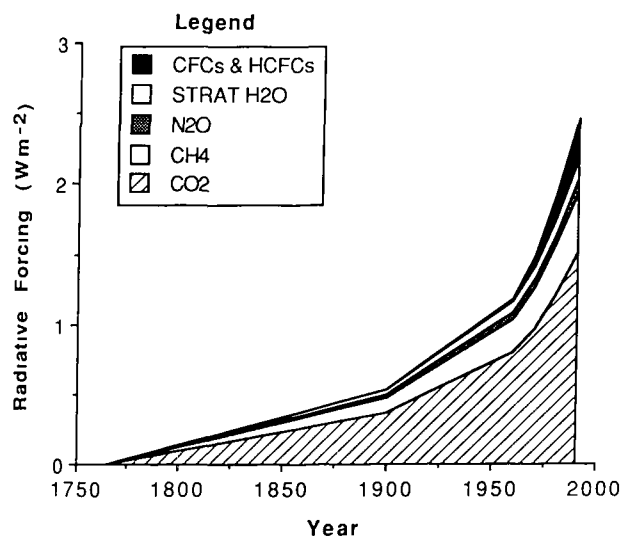
Table 2.6 gives the contributions to the forcing for a number of periods. This is shown diagrammatically in Figure 2.2 as the change in total forcing from 1765 concentrations, it is shown as a change in forcing per decade in Figure 2.3.

**Table 2.5:** Trace gas concentrations from 1765 to 1990, used to construct Figure 2.2

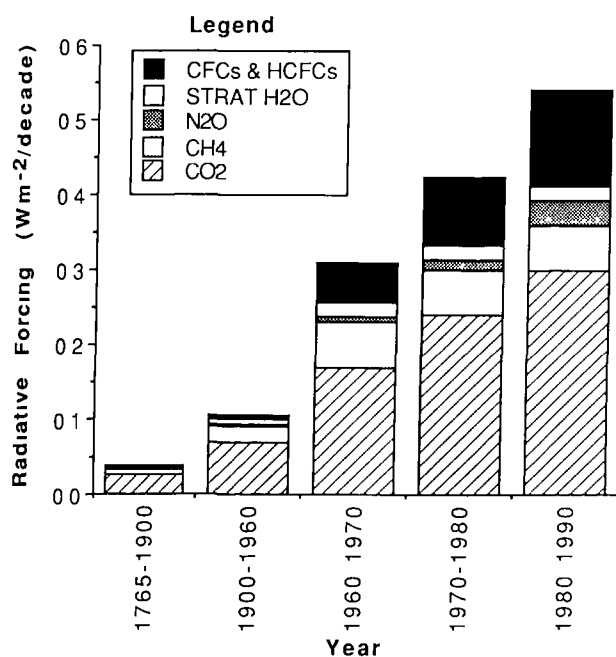
YEAR	CO <sub>2</sub> (ppmv)	CH <sub>4</sub> (ppbv)	N <sub>2</sub> O (ppbv)	CFC-11 (ppbv)	CFC-12 (ppbv)
1765	279.00	790.0	285.00	0	0
1900	295.72	974.1	292.02	0	0
1960	316.24	1272.0	296.62	0.0175	0.0303
1970	324.76	1420.9	298.82	0.0700	0.1211
1980	337.32	1569.0	302.62	0.1575	0.2725
1990	353.93	1717.0	309.68	0.2800	0.4844

**Table 2.6:** Forcing in  $Wm^{-2}$  due to changes in trace gas concentrations in Table 2.5. All values are for changes in forcing from 1765 concentrations. The change due to stratospheric water vapour is an indirect effect of changes in methane concentration (see text)

YEAR	SUM	CO <sub>2</sub>	CH <sub>4</sub> direct	Strat H <sub>2</sub> O	N <sub>2</sub> O	CFC-11	CFC-12	Other CFCs
1765-1900	0.53	0.37	0.10	0.034	0.027	0.0	0.0	0.0
1765-1960	1.17	0.79	0.24	0.082	0.045	0.004	0.008	0.005
1765-1970	1.48	0.96	0.30	0.10	0.054	0.014	0.034	0.021
1765-1980	1.91	1.20	0.36	0.12	0.068	0.035	0.076	0.048
1765-1990	2.45	1.50	0.42	0.14	0.10	0.062	0.14	0.085



**Figure 2.2:** Changes in radiative forcing ( $\text{Wm}^{-2}$ ) due to increases in greenhouse gas concentrations between 1765 and 1990. Values are changes in forcing from 1765 concentrations.



**Figure 2.3:** Decadal contributions to radiative forcing ( $\text{Wm}^{-2}$ ) due to increases in greenhouse gas concentrations for periods between 1765 and 1990. The changes for the periods 1765-1900 and 1900-1960 are the total changes during these periods divided by the number of decades.

Changes in halocarbons other than CFC-11 and CFC-12 have been accounted for by using concentration changes from Section 1 and the forcing versus-concentration changes given in Tables 2.2 and 2.4. It is found that they contribute an extra 43% of the sum of the forcing from CFC-11 and CFC-12, most of this contribution results from

changes in HCFC-22, CFC-113, carbon tetrachloride and methyl chloroform. This is in reasonable agreement with Hansen et al. (1989) who, using less recent spectroscopic data, find these halocarbons contribute an extra 60% of the combined CFC-11 and CFC-12 forcing.

For the period 1765 to 1990,  $\text{CO}_2$  has contributed about 61% of the forcing, methane 17% plus 6% from stratospheric water vapour,  $\text{N}_2\text{O}$  4% and the CFCs 12%. For the decade 1980-1990, about 56% of the forcing has been due to changes in  $\text{CO}_2$ , 11% due to the direct effects of  $\text{CH}_4$  and 4% via stratospheric water vapour, 6% from  $\text{N}_2\text{O}$  and 24% from the CFCs.

As discussed in Section 1, the distribution of tropospheric ozone has almost certainly changed over this period, with a possible impact on radiative forcing. Difficulties in assessing the global changes in ozone, and in calculating the resultant radiative forcing, prevent a detailed assessment of the effect. Estimates of tropospheric ozone change driven by changing methane and  $\text{NO}_x$  emissions are highly model dependent, partly because of the inherent spatial averages used in current two-dimensional models. Estimates of changes in tropospheric ozone from pre-industrial values (e.g. Hough and Derwent 1990) and simplified estimates of the radiative forcing (Table 2.2) suggest that tropospheric ozone may have contributed about 10% of the total forcing due to greenhouse gases since pre-industrial times.

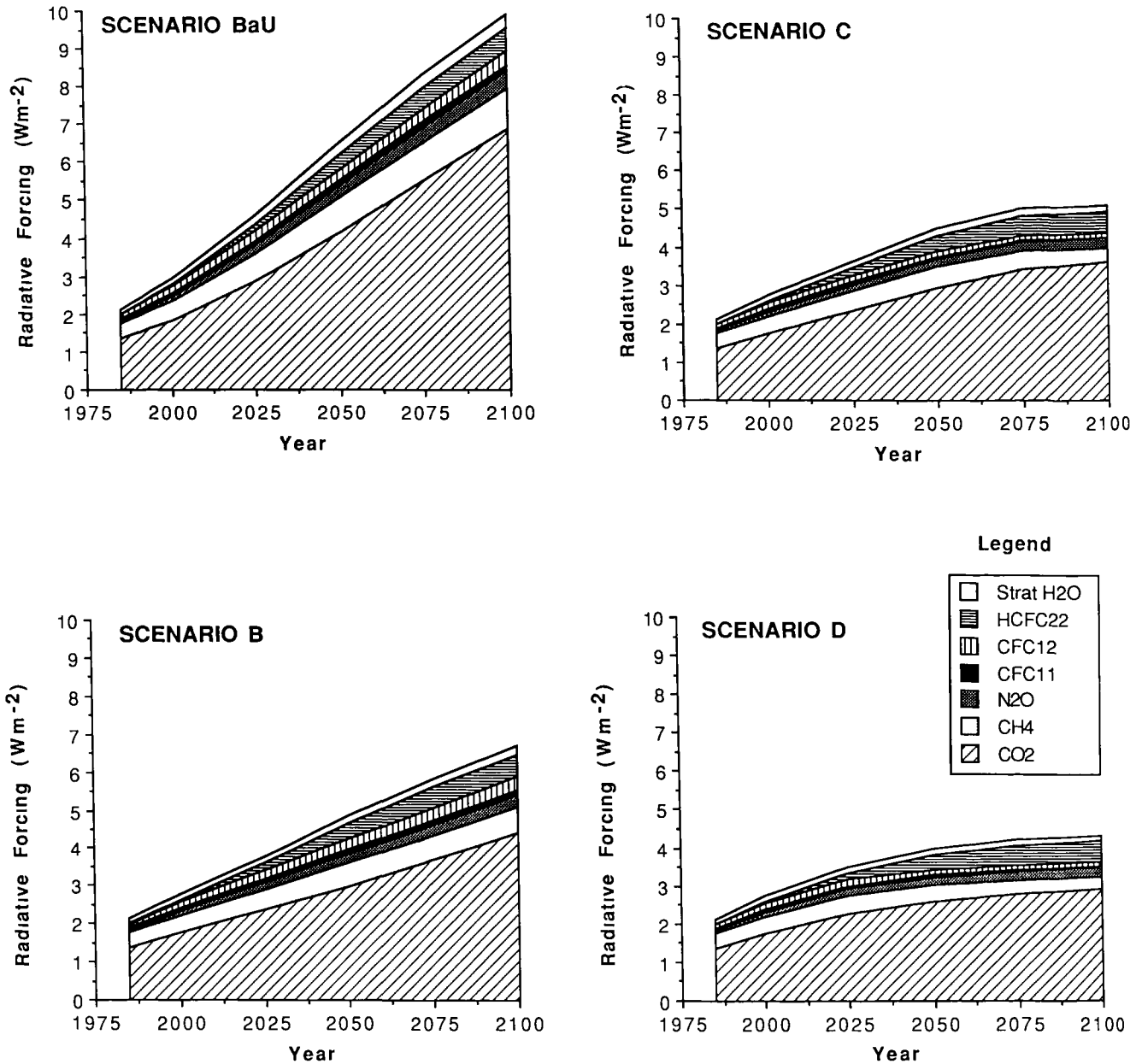
Decreases in lower stratospheric ozone, particularly since the mid-70s, may have led to a decreased radiative forcing; this may have compensated for the effects of tropospheric ozone (Hansen et al., 1989; Lacis et al., 1990). This compensation should be considered as largely fortuitous, as the mechanisms influencing ozone concentrations in the troposphere and stratosphere are somewhat different.

### 2.2.6 Calculations of Future Forcing

Using the radiative forcing expressions described in Section 2.2.4, and the four scenarios developed by Working Group III, possible changes in radiative forcing over the next century can be calculated. The four scenarios are intended to provide insight into policy analysis for a range of potential changes in concentrations. Scenario A is a 'Business as Usual' case, whilst Scenarios B, C and D represent cases of reduced emissions. These four scenarios are considered in more detail in the Appendix 1. As in the previous section, the indirect effect of methane on forcing via stratospheric water vapour changes is included, whilst the effects of possible changes in ozone are neglected.

It must be stressed here that the gas referred to as HCFC-22 as given in the scenarios is used as a surrogate for all the CFC substitutes. Since all HCFCs and HFCs are of similar radiative strength on a molecule-per-molecule basis (see Table 2.4) the error from this source in using HCFC-22 as





**Figure 2.4:** Possible future changes in radiative forcing ( $\text{Wm}^{-2}$ ) due to increases in greenhouse gas concentrations between 1985 and 2100 using the four policy scenarios given in the Appendix 1. Values are changes in forcing from 1765 concentrations.

a proxy for the other gases will be small. However, since the concentrations, as specified in the scenarios, were calculated assuming the HCFC-22 lifetime and molecular weight, considerable errors in the forcing may result from errors in the concentrations. Since some of the CFC substitutes have a longer lifetime than HCFC-22, and some shorter, it is not possible to calculate the sign of the error without knowing the precise mix of substitutes used.

Figure 2.4 shows the radiative forcing change (from pre-industrial) for each gas from the four scenarios; the results are tabulated in Table 2.7.

For these scenarios,  $\text{CO}_2$  remains the dominant contributor to change throughout the period. In the

Business as Usual Scenario, for example, its contribution to the change always exceeds 60%. For the scenarios chosen for this analysis, the contribution of HCFC-22 becomes significant in the next century. It is contributing 11% of the 25-year forcing change between 2025-2050 in the Business-as-Usual Scenario and 18% in Scenario B.

Since the concentration of chlorine can be anticipated to increase in the stratosphere for at least the next decade (Section 1.6.2; see also Prather and Watson, 1990), further decreases in stratospheric ozone can be anticipated. Decreases in upper stratospheric ozone will lead to a small warming effect; decreases in the lower stratosphere would cause a cooling effect. A 1% loss in ozone in the lower

**Table 2 7:** Changes in radiative forcing in  $Wm^2$  for the 4 policy scenarios. The change due to stratospheric water vapour is an indirect effect of changes in methane concentration (see text). All values are changes in forcing from 1765 concentrations.

**SCENARIO A (Business-as-Usual)**

YEAR	SUM	CO <sub>2</sub>	CH <sub>4</sub> direct	Strat H <sub>2</sub> O	N <sub>2</sub> O	CFC-11	CFC-12	HCFC- 22
1765 2000	2.95	1.85	0.51	0.18	0.12	0.08	0.17	0.04
1765 2025	4.59	2.88	0.72	0.25	0.21	0.11	0.25	0.17
1765 2050	6.49	4.15	0.90	0.31	0.31	0.12	0.30	0.39
1765 2075	8.28	5.49	1.02	0.35	0.40	0.13	0.35	0.55
1765 2100	9.90	6.84	1.09	0.38	0.47	0.14	0.39	0.59

**SCENARIO B (Low Emissions)**

YEAR	SUM	CO <sub>2</sub>	CH <sub>4</sub> direct	Strat H <sub>2</sub> O	N <sub>2</sub> O	CFC-11	CFC-12	HCFC- 22
1765 2000	2.77	1.75	0.45	0.16	0.11	0.08	0.17	0.04
1765 2025	3.80	2.35	0.56	0.19	0.18	0.10	0.24	0.17
1765 2050	4.87	2.97	0.65	0.22	0.23	0.11	0.29	0.39
1765 2075	5.84	3.69	0.66	0.23	0.28	0.12	0.33	0.53
1765 2100	6.68	4.43	0.66	0.23	0.33	0.12	0.36	0.56

**SCENARIO C (Control Policies)**

YEAR	SUM	CO <sub>2</sub>	CH <sub>4</sub> direct	Strat H <sub>2</sub> O	N <sub>2</sub> O	CFC-11	CFC-12	HCFC- 22
1765 2000	2.74	1.75	0.44	0.15	0.11	0.08	0.17	0.05
1765 2025	3.63	2.34	0.51	0.17	0.17	0.07	0.17	0.20
1765 2050	4.49	2.96	0.53	0.18	0.22	0.05	0.14	0.41
1765 2075	5.00	3.42	0.47	0.16	0.25	0.03	0.12	0.55
1765 2100	5.07	3.62	0.37	0.13	0.27	0.02	0.10	0.57

**SCENARIO D (Accelerated Policies)**

YEAR	SUM	CO <sub>2</sub>	CH <sub>4</sub> direct	Strat H <sub>2</sub> O	N <sub>2</sub> O	CFC-11	CFC-12	HCFC- 22
1765 2000	2.74	1.75	0.44	0.15	0.11	0.08	0.17	0.04
1765 2025	3.52	2.29	0.47	0.16	0.17	0.07	0.17	0.20
1765 2050	3.99	2.60	0.43	0.15	0.21	0.05	0.14	0.40
1765 2075	4.22	2.77	0.39	0.13	0.24	0.03	0.12	0.53
1765 2100	4.30	2.90	0.34	0.12	0.26	0.02	0.10	0.56

stratosphere would cause a change of about  $0.05 Wm^{-2}$  so that changes could be significant on a decadal time-scale. Possible decreases in chlorine content as a result of international agreements (Prather and Watson 1990) would

be expected to lead to a slow recovery of stratospheric ozone over many decades, which would then result in a small positive forcing over the period of that recovery.

### 2.2.7 A Global Warming Potential Concept for Trace Gases

In considering the policy options for dealing with greenhouse gases, it is necessary to have a simple means of describing the relative abilities of emissions of each greenhouse gas to affect radiative forcing and hence climate. A useful approach could be to express any estimates relative to the trace gas of primary concern, namely carbon dioxide. It would follow on from the concept of relative Ozone Depletion Potential (ODP) which has become an integral part of the Montreal Protocol and other national and international agreements for controlling emissions of halocarbons (e.g. UNEP 1989). The long lifetime of some greenhouse gases implies some commitment to possible climate impacts for decades or centuries to come, and hence the inclusion of potential in the formulation of the concept.

Estimates of the relative greenhouse forcing based on atmospheric concentrations have been detailed in Section 2.2.3. These are relatively straightforward to evaluate. Relative forcings based on emissions are of much greater intrinsic interest to policy makers but require a careful consideration of the radiative properties of the gases, their lifetimes and their indirect effects on greenhouse gases. Wuebbles (1989) has reviewed various approaches to the design of relative forcings based on emissions using past and current trends in global emissions and concentrations.

It must be stressed that there is no universally accepted methodology for combining all the relevant factors into a single global warming potential for greenhouse gas emissions. In fact there may be no single approach which will represent all the needs of policy makers. A simple approach has been adopted here to illustrate the difficulties inherent in the concept, to illustrate the importance of some of the current gaps in understanding and to demonstrate the current range of uncertainties. However, because of the importance of greenhouse warming potentials, a preliminary evaluation is made.

**The Global Warming Potential (GWP)** of the emissions of a greenhouse gas, as employed in this report, is the time integrated commitment to climate forcing from the instantaneous release of 1 kg of a trace gas expressed relative to that from 1 kg of carbon dioxide

$$\text{GWP} = \frac{\int_0^n a_1 c_1 dt}{\int_0^n a_{\text{CO}_2} c_{\text{CO}_2} dt}$$

where  $a_1$  is the instantaneous radiative forcing due to a unit increase in the concentration of trace gas,  $i$ ,  $c_1$  is concentration of the trace gas,  $i$ , remaining at time,  $t$ , after its release and  $n$  is the number of years over which the calculation is performed. The corresponding values for carbon dioxide are in the denominator.

Fisher et al. (1990) have used a similar analysis to derive a global warming potential for halocarbons taken relative to CFC-11. In their work it is implicitly assumed that the integration time is out to infinity.

Early attempts at defining a concept of global warming potentials (Lashof and Ahuja 1990, Rodhe, 1990, Derwent 1990) are based on the instantaneous emissions into the atmosphere of a quantity of a particular trace gas. The trace gas concentration then declines with time and whilst it is present in the atmosphere it generates a greenhouse warming. If its decline is due to atmospheric chemistry processes then the products of these reactions may generate an additional greenhouse warming. A realistic emissions scenario can be thought of as due to a large number of instantaneous releases of different magnitudes over an extended time period and some emission abatement scenarios can be evaluated using this concept.

Particular problems associated with evaluating the GWP are

- the estimation of atmospheric lifetimes of gases (and in particular  $\text{CO}_2$ ), and the variation of that lifetime in the future,
- the dependence of the radiative forcing of a gas on its concentration and the concentration of other gases with spectrally overlapping absorption bands
- the calculation of the indirect effects of the emitted gases and the subsequent radiative effects of these indirect greenhouse gases (ozone poses a particular problem),
- the specification of the most appropriate time period over which to perform the integration

The full resolution of the above problems must await further research. The assumptions made in the present assessment are described below.

For some environmental impacts, it is important to evaluate the cumulative greenhouse warming over an extended period after the instantaneous release of the trace gas. For the evaluation of sea-level rise, the commitment to greenhouse warming over a 100 year or longer time horizon may be appropriate. For the evaluation of short term effects, a time horizon of a few decades could be taken, for example, model studies show that continental areas are able to respond rapidly to radiative forcing (see e.g., Section 6) so that the relative effects of emissions on such timescales are relevant to predictions of near-term climate change. This consideration alone dramatically changes the emphasis between the different greenhouse

gases, depending on their persistence in the atmosphere. For this reason, global warming potentials in Table 2.8 have been evaluated over 20, 100 and 500 years. These three different time horizons are presented as candidates for discussion and should not be considered as having any special significance.

The figures presented in Table 2.8 should be considered preliminary only. Considerable uncertainty exists as to the lifetimes of methane and many of the halocarbons, due to difficulties in modelling the chemistry of the troposphere. The specification of a single lifetime for carbon dioxide also presents difficulties, this is an approximation of the actual lifetime due to the transfer of CO<sub>2</sub> amongst the different reservoirs. The detailed time behaviour of a pulse of carbon dioxide added to the atmosphere has been described using an ocean-atmosphere-biosphere carbon dioxide model (Siegenthaler, 1983). The added carbon dioxide declines in a markedly non-exponential manner: there is an initial fast decline over the first 10 year period, followed by a more gradual decline over the next 100 years and a rather slow decline over the thousand year time-scale. The time period for the first half-life is typically around 50 years for the second, about 250 years (see Section 1.2.1 for details). A single lifetime figure defined by the decline to 1/e is about 120 years. Indeed the uncertainties associated with specifying the lifetime of CO<sub>2</sub> means that presentation of the GWP relative to CO<sub>2</sub> may not be the ideal choice, relative GWPs of gases other than CO<sub>2</sub> to each other are not affected by this uncertainty.

In performing the integration of greenhouse impacts into the future a number of simplifications have been made. The neglect of the dependence of the radiative term on the trace gas concentration implies small trace gas concentration changes. Further, the overlap of the infrared absorption bands of methane and nitrous oxide may be significant and this restricts the application of the GWP to small perturbations around present day concentrations.

An assumption implicit in this simple approach is that the atmospheric lifetimes of the trace gases remain constant over the integration time horizon. This is likely to be a poor assumption for many trace gases for a variety of different reasons. For those trace gases which are removed by tropospheric OH radicals, a significant change in lifetime could be anticipated in the future, depending on the impact of human activities on methane, carbon monoxide and oxides of nitrogen emissions. For some scenarios, as much as a 50% increase in methane and HCFC 22 lifetimes has been estimated. Such increases in lifetime have a dramatic influence on the global warming potentials in Table 2.8, integrated over the longer time horizons. Much more work needs to be done to determine global warming potentials which will properly account for the processes affecting atmospheric composition and for the possible non-linear

feedbacks influencing the impacts of trace gases on climate.

It is recognised that the emissions of a number of trace gases, including NO<sub>x</sub>, carbon monoxide, methane and other hydrocarbons, have the potential to influence the distribution of tropospheric ozone. It is not straightforward to estimate the greenhouse warming potential of these indirect effects because changes in tropospheric ozone depend, in a complex and non-linear manner on the concentrations of a range of species. The limited spatial resolution in current tropospheric chemistry models means that estimates of increased tropospheric ozone production are highly model-dependent. Furthermore, the radiative impacts of tropospheric ozone changes depend markedly on their spatial distribution. As a result, the GWP values for the secondary greenhouse gases have been provided as first order estimates only, using results from a tropospheric two-dimensional model of global atmospheric chemistry (Hough and Derwent, 1990) and the radiative forcing given in Table 2.2 (see Derwent (1990) for further details). Evaluation of the radiative forcing resulting from changes in concentrations of stratospheric ozone (as a result of CFC, N<sub>2</sub>O, and CH<sub>4</sub> emissions) have not been included due to insufficient time to undertake the analysis this requires.

Bearing in mind the uncertainties inherent in Table 2.8, a number of important points are raised by the results. Firstly, over a twenty year period a kilogram of all the proposed CFC substitutes, with the exception of the relatively short lived HCFC-123 and HFC-152a, cause more than a three order of magnitude greater warming than 1 kg of CO<sub>2</sub>. However, for a number of these gases (but *not* the five CFCs themselves) the global warming potential reduces markedly as the integration time is increased, this implies that over the long term, the replacement compounds should have a much lower global warming effect than the CFCs they replace, for the same levels of emissions. In addition, the shorter lifetimes imply that abrupt changes in total emissions would impact on the actual global warming relatively quickly. A further important point is that in terms of radiative forcing over the short-term the effect of the CFC substitutes is considerably greater than indicated by the halocarbon global warming potential (GWP) of Fisher et al (1990). For example, over a 20 year period, the effect of 1 kg emission of HCFC 22 contributes only slightly less to the radiative forcing than the same amount of CFC-11, even though its 'infinite' GWP is about 0.35. This is because, on a kg-per-kg basis, HCFC 22 is a stronger greenhouse gas than CFC-11 (Table 2.4).

The indirect greenhouse warmings listed in Table 2.8 are potentially very significant. The production of CO<sub>2</sub> stratospheric water vapour and tropospheric ozone as a result of emissions of methane leads to an indirect effect

**Table 2.8:** Global warming potentials following the instantaneous injection of 1 kg of each trace gas, relative to carbon dioxide. A specific example of an application of these potentials is given in Table 2.9

Trace Gas	Estimated Lifetime, years	Global Warming Potential		
		Integration Time Horizon, Years		
		20	100	500
Carbon Dioxide	*	1	1	1
Methane - inc indirect	10	63	21	9
Nitrous Oxide	150	270	290	190
CFC-11	60	4500	3500	1500
CFC-12	130	7100	7300	4500
HCFC-22	15	4100	1500	510
CFC-113	90	4500	4200	2100
CFC-114	200	6000	6900	5500
CFC-115	400	5500	6900	7400
HCFC 123	1.6	310	85	29
HCFC-124	6.6	1500	430	150
HFC-125	28	4700	2500	860
HFC-134a	16	3200	1200	420
HCFC-141b	8	1500	440	150
HCFC-142b	19	3700	1600	540
HFC-143a	41	4500	2900	1000
HFC 152a	1.7	510	140	47
CCl <sub>4</sub>	50	1900	1300	460
CH <sub>3</sub> CCl <sub>3</sub>	6	350	100	34
CF <sub>3</sub> Br	110	5800	5800	3200
<b>INDIRECT EFFECTS</b>				
Source Gas	Greenhouse Gas Affected			
CH <sub>4</sub>	Tropospheric O <sub>3</sub>	24	8	3
CH <sub>4</sub>	CO <sub>2</sub>	3	3	3
CH <sub>4</sub>	Stratospheric H <sub>2</sub> O	10	4	1
CO	Tropospheric O <sub>3</sub>	5	1	0
CO	CO <sub>2</sub>	2	2	2
NO <sub>x</sub>	Tropospheric O <sub>3</sub>	150	40	14
NMHC	Tropospheric O <sub>3</sub>	28	8	3
NMHC	CO <sub>2</sub>	3	3	3

CFCs and other gases do not include effect through depletion of stratospheric ozone

Changes in lifetime and variations of radiative forcing with concentration are neglected. The effects of N<sub>2</sub>O forcing due to changes in CH<sub>4</sub> (because of overlapping absorption), and vice versa, are neglected

\* The persistence of carbon dioxide has been estimated by explicitly integrating the box-diffusion model of Siegenthaler (1983) - an approximate lifetime is 120 years

almost as large as the direct effect for integration times of a century or longer. The potential for emissions of gases, such as CO, NO<sub>x</sub> and the non-methane hydrocarbons, to contribute indirectly to global warming is also significant. It must be stressed that these indirect effects are highly

model dependent and they will need further revision and evaluation. An example of uncertainty concerns the impact of NO<sub>x</sub> emissions, these emissions generate OH which leads to increased destruction of gases such as methane (e.g., Thompson et al., 1989). This would constitute a

**Table 2.9:** Example of use of Global Warming Potentials. The table shows the integrated effects over a 100 year time horizon of total emissions in 1990 given as a fraction of the total effect

Trace Gas	Current Man Made Emissions Tg yr <sup>-1</sup>	Proportion of total effects %
CO <sub>2</sub>	26000	61
CH <sub>4</sub>	300	15
N <sub>2</sub> O	6	4
CFC-11	0.3	2
CFC-12	0.4	7
HCFC-22	0.1	0.4
CFC-113	0.15	1.5
CFC-114	0.015	0.2
CFC-115	0.005	0.1
CCl <sub>4</sub>	0.09	0.3
CH <sub>3</sub> CCl <sub>3</sub>	0.81	0.2
CO	200	1
NO <sub>x</sub>	66	6
NMHCs	20	0.5

Carbon dioxide emissions given on CO<sub>2</sub> basis, equivalent to 7 GtC yr<sup>-1</sup>. Nitrous oxide emissions given on N<sub>2</sub>O basis, equivalent to 4 MtN yr<sup>-1</sup>. NO<sub>x</sub> emissions given on NO<sub>2</sub> basis equivalent to 20 MtN yr<sup>-1</sup>.

negative indirect effect of NO<sub>x</sub> emissions which would oppose the forcing due to increased tropospheric ozone formation

As an example of the use of the Global Warming Potentials, Table 2.9 shows the integrated effects over a 100 year time horizon for the estimated human-related greenhouse gas emissions in 1990. The derived cumulative effects, derived by multiplying the appropriate GWP by the 1990 emissions rate, indicates that CO<sub>2</sub> will account for 61% of the radiative forcing over this time period. Emissions of NO<sub>x</sub>, whose effect is entirely indirect, is calculated to contribute 6% to the total forcing.

## 2.3 Other Radiative Forcing Agents

### 2.3.1 Solar Radiation

The Sun is the primary source of energy for the Earth's climate system. Variations in the amount of solar radiation received by the Earth can affect our climate. There are two distinct sources of this variability. The first, which acts

with greatest impact on time-scales of 10,000 to 100,000 years is caused by changes in the Sun-Earth orbital parameters. The second comes from physical changes on the Sun itself, such changes occur on almost all time-scales.

#### 2.3.1.1 Variability due to orbital changes

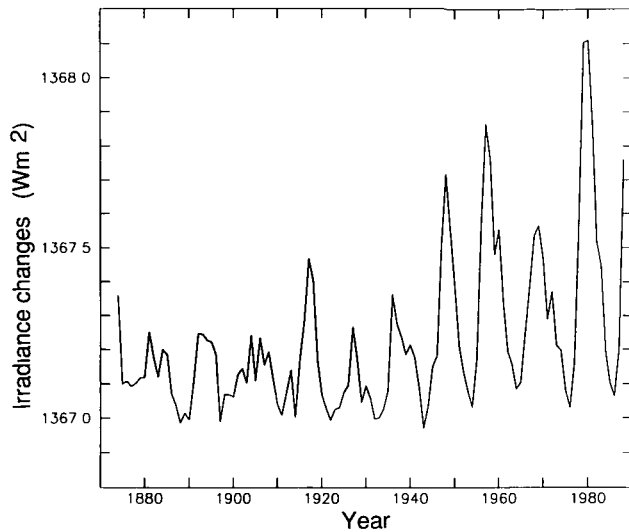
Variations in climate on time-scales ranging from 10,000 to 100,000 years, including the major glacial/interglacial cycles during the Quaternary period, are believed to be initiated by variations in the Earth's orbital parameters which in turn influence the latitudinal and seasonal variation of solar energy received by the Earth (the Milankovitch Effect). Although the covariation of these orbital parameters and the Earth's climate provides a compelling argument in favour of this theory, internal feedback processes have to be invoked to explain the observed climatic variations, in particular the amplitude of the dominating 100,000 year period. One such feedback could be the changes to the carbon cycle and the greenhouse effect of atmospheric CO<sub>2</sub> (see Section 1).

The radiative forcing associated with the Milankovitch Effect can be given for particular latitudes and months to illustrate that the rate of change of forcing is small compared to radiative forcing due to the enhanced greenhouse effect, of course, the climatic impact of the Milankovitch Effect results from the redistribution of solar energy, latitudinally and seasonally, so that a comparison is necessarily rather rough. As an example, in the past 10,000 years, the incident solar radiation at 60°N in July has decreased by about 35 Wm<sup>-2</sup> (e.g., Rind et al., 1989), the average change in one decade is -0.035 Wm<sup>-2</sup>, compared with the estimate, in Section 2.2.5, that the greenhouse forcing over the most recent decade increased by 0.6 Wm<sup>-2</sup> more than 15 times higher than the Milankovitch forcing.

#### 2.3.1.2 Variability due to changes in total solar irradiance

Variations in the short-wave and radio-frequency outputs of the Sun respond to changes in the surface activity of the star and follow in phase with the 11-year sunspot cycle. The greatest changes, in terms of total energy, occur in the short-wave region, and particularly the near ultraviolet. At 0.3 μm, the solar cycle variation is less than 1%, since only about 1% of the Sun's radiation lies at this or shorter wavelengths. Solar-cycle variations in the ultraviolet will by themselves induce variations of no more than 0.01% in total irradiance, although these may be important for atmospheric chemistry in the middle atmosphere.

Of greater potential importance, in terms of direct effects on climate, are changes integrated over all wavelengths: the total solar irradiance or the so-called solar constant. Continuous spaceborne measurements of total irradiance have been made since 1978. These have shown that on time-scales of days to a decade there are irradiance



**Figure 2.5:** Reconstructed solar irradiance ( $\text{Wm}^{-2}$ ) from 1874 to 1988 using the model of Foukal and Lean (1990). The model was calibrated using direct observations of solar irradiance from satellites between 1980 and 1988. Data from J Lean (pers comm). Note that the solar forcing is only 0.175 times the irradiance due to area and albedo effects.

variations that are associated with activity in the Sun's outer layer, the photosphere—specifically, sunspots and bright areas known as faculae. The very high frequency changes are too rapid to affect the climate noticeably. However, there is a lower frequency component that follows the 11-year sunspot cycle which may have a climatic effect. It has been found that the increased irradiance due to faculae more than offsets the decreases due to the cooler sunspots; consequently, high sunspot numbers are associated with high solar output (Foukal and Lean, 1990). Over the period 1980–86, there was a decline in irradiance of about  $1 \text{ Wm}^{-2}$  corresponding to a globally-averaged forcing change at the top of the atmosphere of a little less than  $0.2 \text{ Wm}^{-2}$ . Since then irradiance has increased, following the sunspot cycle (e.g., Willson and Hudson, 1988).

This is comparable with the greenhouse forcing which over the period 1980–86, increased by about  $0.3 \text{ Wm}^{-2}$ . However, over longer periods these solar changes would have contributed only minimally towards offsetting the greenhouse effect on global-mean temperature because of the different time-scales on which the two mechanisms operate. Because of oceanic thermal inertia (see Section 6), and because of the relatively short time scale of the forcing changes associated with the solar cycle, only a small fraction of possible temperature changes due to this source can be realised (Wigley and Raper, 1990). In contrast, the *sustained* nature of the greenhouse forcing allows a much greater fraction of the possible temperature change to be realised, so that the greenhouse forcing dominates.

Because the satellite record of solar irradiance began so recently, we cannot say with absolute certainty what past variations may have been. However, a physically based statistical model has been developed by Foukal and Lean (1990), which attempts to reconstruct the solar-cycle-related changes back more than 100 years (see Figure 2.5). This figure illustrates that the changes from 1980 to 1986 were probably the largest in the past century.

While the model of Foukal and Lean (1990) indicates that the direct effects of solar-cycle-related irradiance changes may have been very small, this does not rule out the possibility of larger, lower-frequency effects. Three possibilities have been hypothesized, they are not supported by direct observational evidence of solar irradiance variations, and their magnitudes are derived by assuming that observed or inferred temperature variations are responses to solar forcing. The first idea is that on the time scale of about a century, some underlying variation exists that parallels the envelope of sunspot activity, i.e. the smooth curve joining the peaks of successive sunspot maxima (Eddy, 1977; Reid, 1987). The envelope curve shows a quasi-cyclic behaviour with period about 80–90 years, referred to as the Gleissberg cycle (e.g., Gilliland, 1982; Gilliland and Schneider, 1984).

There is no reason why one should expect the envelope curve to be related to solar irradiance variations beyond those associated with the Foukal-Lean mechanism. Reid's study appears to have been spurred by the visual similarity between the Folland et al. (1984) global marine temperature curve and the envelope curve. This similarity is less apparent when more recently compiled temperatures are considered (see e.g., Section 7) and is much less apparent in the Southern Hemisphere than in the Northern. With no way to estimate the range of irradiance variation *a priori*, Reid tuned this to obtain a best match between modelled and observed temperatures. Assuming solar change as the sole forcing mechanism, the implied decadal time-scale irradiance range is about 0.6%, or  $1.5 \text{ Wm}^{-2}$  at the top of the atmosphere, for an assumed climate sensitivity of  $2.5^\circ\text{C}$  for a  $\text{CO}_2$  doubling. This value is about thirty times that inferred by direct satellite data.

Reid emphasizes that his work is mainly an exercise in curve-fitting, so that the results should be used with extreme caution. Nevertheless, it has been taken seriously by the Marshall Institute (1989) so a brief analysis is in order. Kelly and Wigley (1990) have performed a similar analysis to Reid's, incorporating a greenhouse forcing history (Section 2.2.5) and using more recent temperature compilations (Section 7). The amplitude  $A$  of the radiative forcing due to solar variability (which is tied to sunspot number) is evaluated so as to give the best agreement between observed and modelled temperatures between 1861 and 1989. The value of  $A$  which gives the best fit is found to depend critically on the assumed climate

sensitivity. For values of equilibrium change due to doubled CO<sub>2</sub> (see Section 5.2.1) of greater than 2°C, it is found that the best fit is obtained if there is a negative correlation between solar output and sunspot number (which contradicts recent observations). At the lower end of the range of climate sensitivity suggested in Section 5.2.1 (1.5°C) the best fit is obtained for a value of  $A$  about one fifth that derived by Reid. However, even for this value the percentage variance explained is only marginally better when the solar and greenhouse effects are considered together than when greenhouse forcing is considered alone. This analysis provides no evidence for low-frequency irradiance variations larger than the small changes that have been directly inferred from satellite based irradiance observations.

The second suggested solar effect makes use of the relationship between solar radius variations and irradiance changes. Radius variations have been observed over the past few centuries but whether these could have significant irradiance changes associated with them is unknown. The proportionality constant relating radius and irradiance changes is so uncertain that it could imply an entirely negligible or a quite noticeable irradiance variation (Gilliland 1982; Wigley 1988). Gilliland (1982) therefore attempted to estimate the solar effect empirically by comparing modelled and observed data. Gilliland concluded that solar induced quasi-cyclic temperature changes (~80 year cycle) with range about 0.2°C might exist but to obtain a reasonable fit he had to invoke a phase lag between radius and irradiance changes. Most theories relating radius and irradiance changes do not allow such a phase lag although an exception has been noted by Wigley (1988). While the physical basis for the radius effect is at least reasonable these results are far from being convincing in a statistical sense as Gilliland himself noted. Nevertheless we cannot completely rule out the possibility of solar forcing changes related to radius variations on an 80 year time-scale causing global mean temperature fluctuations with a range of up to 0.2°C. Hansen and Lacis (1990) regard about 0.8 Wm<sup>-2</sup> as a probable upper limit for the change in forcing due to variations in solar output over such periods.

The third suggested solar effect is that related to the minima in sunspot activity such as the Maunder Minimum for which the associated changes in atmospheric radiocarbon content are used as a proxy. These ideas were revived by Eddy (1977). The hypothesis has some credence in that the sunspot minima are manifestations of solar change (although irradiance changes associated with them would be only a few tenths Wm<sup>-2</sup> based on the Foukal-Lean model) as are radiocarbon fluctuations. But neither is direct evidence of solar irradiance changes. Indirect evidence of irradiance changes comes from the climate record specifically the observation that during the

Holocene the timing of the neoglacial (i.e. Little Ice Age type) events show some correspondence with times of anomalous atmospheric radiocarbon content. Wigley (1988) and Wigley and Kelly (1990) found the correlation over a 10 000 year period to be statistically significant but far from convincing. Nevertheless, if one accepts its reality, the magnitude of the solar forcing changes required to cause the observed neoglacial events can be shown to have been up to 1.3 Wm<sup>-2</sup> at the top of the atmosphere, averaged over 100 200 years. These results have also been used by the Marshall Institute (1989) who suggest that another Little Ice Age is imminent and that this may substantially offset any future greenhouse-gas-induced warming. While one might expect such an event to occur some time in the future the timing cannot be predicted. Further the 1.3 Wm<sup>-2</sup> solar change (which is an upper limit) is small compared with greenhouse forcing and even if such a change occurred over the next few decades, it would be swamped by the enhanced greenhouse effect.

### 2.3.2 Direct Aerosol Effects

The impact of aerosol particles, i.e. solid or liquid particles in the size range 0.001-10 μm radius, on the radiation budget of the Earth-atmosphere system is manifold, either *directly* through scattering and absorption in the solar and thermal infrared spectral ranges or *indirectly* by the modification of the microphysical properties of clouds which affects their radiative properties. There is no doubt that aerosol particles influence the Earth's climate. However their influence is far more difficult to assess than that of the trace gases because they constitute their own class of substances with different size distributions, shape, chemical compositions and optical properties and because their concentrations vary by orders of magnitude in space and time and because observations of their temporal and spatial variation are poor (Section 1).

It is not easy to determine the sign of changes in the planetary radiation budget due to aerosols. Depending on absorption-to-backscattering ratio, surface albedo, total aerosol optical depth and solar elevation - if ordered approximately according to importance - additional aerosol particles may either increase or decrease local planetary albedo (e.g. Coakley and Chylek 1975; Grassl and Newiger 1982). A given aerosol load may increase the planetary albedo above an ocean surface and decrease it above a sand desert. The effect of aerosol particles on terrestrial radiation cannot be neglected, in conditions where the albedo change is small, the added greenhouse effect can dominate (Grassl 1988).

While it is easy to demonstrate that aerosol particles measurably reduce solar irradiance in industrial regions the lack of data and inadequate spatial coverage preclude extending this demonstration to larger spatial scales. For example Ball and Robinson (1982) have shown for the



eastern U.S. an average annual depletion of solar irradiance of 7.5% at the surface. Some of this depleted radiation will, however, have been absorbed within the troposphere, so that the perturbation to the net flux at the tropopause will be somewhat less and the impact on the thermal infrared is not quantified. Most of this perturbation is anthropogenic. The depletion is regionally very significant, for example, for a daily mean surface irradiance of  $200 \text{ Wm}^{-2}$ , if about half of the depleted irradiance is lost to space, the change in forcing would be  $7.5 \text{ Wm}^{-2}$ .

Carbon black (soot) plays an especially important role for the local heating rate in the air as it is the only strong absorber in the visible and near infrared spectrum present in aerosol particles. Soot incorporated into cloud particles can also directly affect the radiative properties of clouds by decreasing cloud albedo and hence lead to a positive forcing (e.g., Grassl 1988).

In view of the above uncertainties on the sign, the affected area and the temporal trend of the direct impact of aerosols, we are unable to estimate the change in forcing due to tropospheric aerosols.

Concentrations of stratospheric aerosols may be greatly enhanced over large areas for a few years following large explosive volcanic eruptions although there is no evidence for any secular increase in background aerosol (Section 1).

Major volcanic eruptions can inject gaseous sulphur dioxide and dust, among other chemicals, into the stratosphere. The sulphur dioxide is quickly converted into sulphuric acid aerosols. If present in sufficient quantities in the stratosphere, where the half-life is about 1 year, these aerosols can significantly affect the net radiation balance of the Earth.

These aerosols can drastically reduce (by up to tens of percent) the direct solar beam, although this is, to some extent, compensated by an increase in diffuse radiation, so that decreases in total radiation are smaller (typically 5-10%) (e.g., Spaenkuch, 1978; Coulson, 1988). This decrease in insolation, coupled with the warming due to the thermal infrared effects of the aerosols, leaves only a small deficit in the radiative heating at the surface, for even a major volcanic eruption. Furthermore, volcanic aerosol clouds usually cover only a limited portion of the globe and they exist for a time (1-3 years) that is short compared to the response time of the ocean-atmosphere system (which is of order decades). Thus their climatic effects should be relatively short-lived. Because the size distribution and the optical properties of the particles are very important in determining whether the Earth's surface warms or cools, theoretical estimates of their effect on the surface climate are strongly dependent on the assumptions made about the aerosols (e.g., Mass and Portman (1989) and references therein).

A number of empirical studies have been carried out to detect the impact of volcanic eruptions on surface

temperatures over the last 100 years or more (e.g., Bradley, 1988; Mass and Portman, 1989). Generally these studies have concluded that major volcanic events, of which there were only about 5 during the past century, may cause a global-mean cooling of 0.1 to 0.2°C for a one to two year period after the event. A direct calculation of the radiative impact of a major volcanic eruption (Ramanathan, 1988) shows that the decadal radiative forcing may be 0.2 - 0.4  $\text{Wm}^{-2}$ , indicating that they can have a significant climatic impact on decadal time-scales.

There have also been claims of longer time-scale effects. For example, Hammer et al. (1980) and Porter (1987) have claimed that the climate fluctuations of the last millenium, including events like the Little Ice Age, were due largely to variations in explosive volcanic activity, and various authors have suggested that decadal time-scale trends in the twentieth century were strongly influenced by the changing frequencies of large eruptions (SCOPE, 1986). These claims are highly contentious and generally based on debatable evidence. For instance, a major problem in such studies is that there is no agreed record of past volcanic forcing - alternative records published in the literature correlate poorly. In consequence, the statistical evidence for a low frequency volcanic effect is poor (Wigley et al 1986) but not negligible (Schonwiese, 1988); since the lifetime of the aerosols in the stratosphere is only a few years, such an effect would require frequent explosive eruptions to cause long time-scale fluctuations in aerosol loading.

In summary, there is little doubt that major volcanic eruptions contribute to the interannual variability of the global temperature record. There is no convincing evidence, however, of longer time-scale effects. In the future, the effects of volcanic eruptions will continue to impose small year-to-year fluctuations on the global mean temperature. Furthermore, a period of sustained intense volcanic activity could partially offset or delay the effects of warming due to increased concentrations of greenhouse gases. However, such a period would be plainly evident and readily allowed for in any contemporary assessment of the progress of the greenhouse warming.

### 2.3.3 Indirect Aerosol Effects

Cloud droplets form exclusively through condensation of water vapour on cloud condensation nuclei (CCN): i.e., aerosol particles. Therefore, the size, number and the chemical composition of aerosol particles, as well as updraughts, determine the number of cloud droplets. As a consequence, continental clouds, especially over populated regions, have a higher droplet concentration (by a factor of order 10) than those in remote marine areas. Clouds with the same vertical extent and liquid water content are calculated to have a higher short-wave albedo over continents than over the oceans (e.g., Twomey, 1977). In

other words, the more polluted an area by aerosol particles the more reflective the clouds. This effect is most pronounced for moderately thick clouds such as marine stratocumulus and stratus clouds which cover about 25% of the Earth's surface. Hence, an increased load of aerosol particles has the potential to increase the albedo of the planet and thus to some extent counteract the enhanced greenhouse effect.

The strongest confirmation of this aerosol/cloud albedo connection stems from observations of clouds in the wake of ship-stack effluents. Ships enhance existing cloud cover (Twomey et al., 1984), and measurably increase the reflectivities (albedo) of clouds in overcast conditions (Coakley et al., 1987). While the *in-situ* observations (Radke et al., 1990) have shown the expected increase in droplet numbers and decrease in droplet sizes for the contaminated clouds, they have also shown an increase in cloud liquid water content (LWC) in contradiction to the suggestion by Twomey et al. (1984) that the changes in the droplet size distribution will leave the LWC nearly unchanged. Albrecht (1989) has suggested that the LWC increase could be due to the suppression of drizzle in the contaminated clouds. An increase of the number of CCN therefore may have an even more complicated influence than has been analysed.

The increase in aerosol sulphate caused by anthropogenic SO<sub>2</sub> emissions (Section 1 Figure 1.16) may have caused an increase in the number of CCN with possible subsequent influence on cloud albedo and climate.

Cess (personal communication) has reported changes in planetary albedo over cloudy skies that are consistent with a larger-scale effect of sulphate emissions. Measurements from the Earth Radiation Budget Experiment satellite instruments indicate, after other factors have been taken into account, that the planetary albedo over low clouds decreases by a few per cent between the western and eastern North Atlantic. The implication is that sulphate emissions from the east coast of North America are affecting cloud albedos downwind. A similar effect can be seen in the North Pacific off the coast of Asia.

There are important gaps in our understanding and too little data, so that a confident assessment of the influence of sulphur emissions on radiative forcing cannot be made. Wigley (1989) has estimated a global-mean forcing change of between -0.25 and -1.25 Wm<sup>-2</sup> from 1900 to 1985 (with all of it actually occurring in the Northern Hemisphere). Deriving a forcing history during this period presents even further difficulties, so that we use, for a typical decadal forcing, the average change of -0.03 to -0.15 Wm<sup>-2</sup> per decade.

Reference to Figure 2.3 shows that this forcing may have contributed significantly to the total forcing, particularly earlier in the century, at these times it may have been of a similar size, but of opposite sign, to the forcing caused by

the enhanced greenhouse effect. Indeed, it has been suggested that the increase in CCN of industrial origin (see Section 1.7.1) might explain why the Northern Hemisphere has not been warming as rapidly as the Southern Hemisphere over the last 50 years. Wigley (1989) estimates that each 0.1°C increase in the twentieth century warming of the Southern Hemisphere relative to the Northern Hemisphere corresponds to a mean forcing differential of around -0.5 Wm<sup>-2</sup>, or a CCN increase of about 10%.

Sulphur emissions are actively being reduced in many countries. Hence, even if some compensation in the total forcing is occurring because of changes in sulphate and greenhouse gases, it is not clear whether that compensation will continue in the future. Because of the limited atmospheric residence time of the sulphur compounds, their possible effects on climate will be reduced as soon as their emissions are decreased. A decrease in sulphur emissions would, via this theory, cause a decrease in cloud albedo. The change in forcing over a decade could then be positive (although the total change from pre-industrial times would remain less than or equal to zero). Hence we are unable to estimate even the sign of future changes in forcing due to this sulphate effect.

A further important point is that even if the cloud albedo increases exactly offset the forcing due to increased concentrations of greenhouse gases, this would not necessarily imply zero climate change. The sulphate effect would tend to act only regionally, whilst the greenhouse forcing is global. Hence regional climate change would still be possible even if the global mean perturbation to the radiation balance were to be zero.

### 2.3.4 Surface Characteristics

The effects of desertification, salinization, temperate and tropical deforestation and urbanization on the surface albedo have been calculated by Sagan et al. (1979). They calculated an absolute change in surface albedo of  $6 \times 10^{-3}$  over the last 1000 years and  $1 \times 10^{-3}$  over as short a time as the last 25 years. Henderson-Sellers and Gornitz (1984) updated these latter calculations to a maximum albedo change over the last 25-30 years of between 3.3 and 6.4  $\times 10^{-4}$ . From Hansen et al. (1988) the radiative forcing (in Wm<sup>-2</sup>) for a change in a land surface albedo is about

$$\Delta F = 43 \Delta x \quad (\Delta x \leq 0.1)$$

where  $\Delta x$  is the change (as a decimal fraction) in the land albedo. (The expression implicitly accounts for the fact that the land surface occupies only 30% of the total surface area of the globe.)

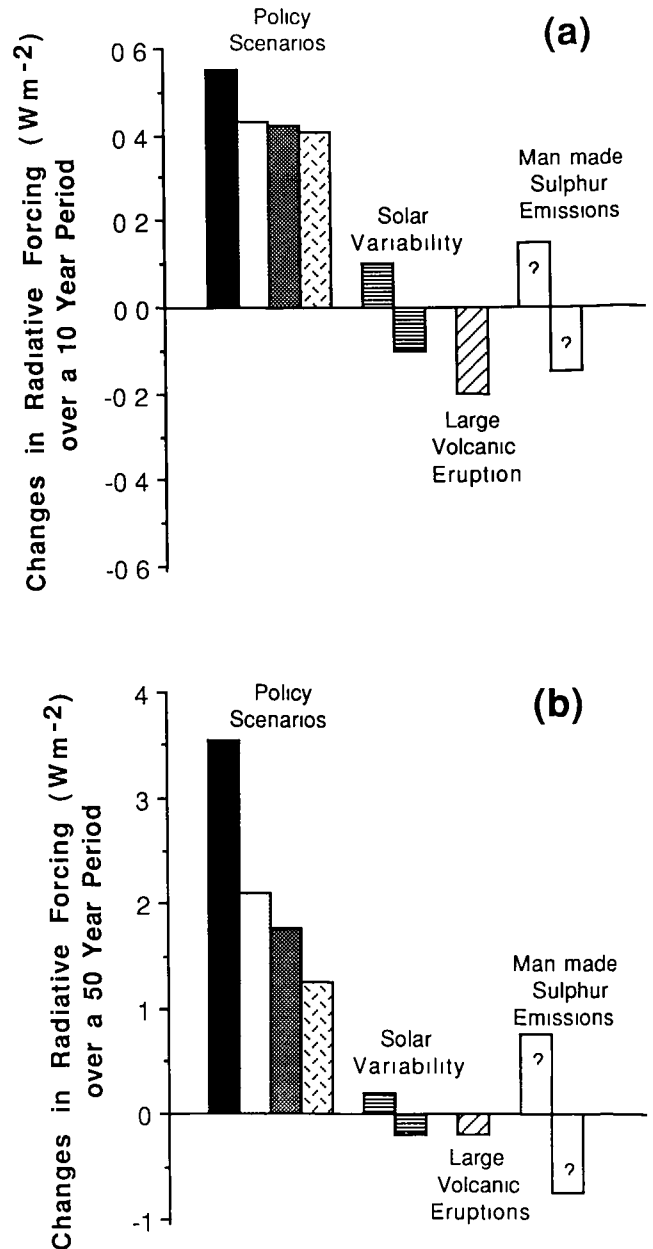
Thus the albedo change over the last few decades will have produced a radiative forcing of 0.03 Wm<sup>-2</sup> at most, i.e. the effects of surface albedo changes on the planetary radiation budget are very small. The effects of changes in surface characteristics on water balance and surface

roughness are likely to be far more important for the regional climate, the changes are discussed in Section 5.6

## 2.4 The Relative Importance of Radiative Forcing Agents in the Future

The analyses of past trends and future projections of the changes in concentrations of greenhouse gases indicate that the radiative forcing from these gases may increase by as much as  $0.4\text{--}0.6\text{ Wm}^{-2}$  per decade over the next several decades. As discussed in Section 2.3, decadal-scale changes in the radiative forcing can also result from other causes. Natural effects on the forcing as a result of solar variability and volcanic eruptions are particularly relevant on decadal timescales. Other potentially important anthropogenic effects may result from increases in the aerosol content of the lower atmosphere, particularly as a result of sulphur emissions. It is important to consider how these additional forcings may modify the atmospheric radiative forcing from that expected from greenhouse gases on both decadal and longer timescales.

Over the period of a decade, the other radiative forcings could extensively modify the expected radiative forcing from greenhouse gases. The additional forcing could either add to, subtract from, or even largely negate the radiative forcing from greenhouse gases, with the effect over any given decade possibly being quite different from that over other decades. Figure 2.6(a) estimates the range of possible effects from solar variability, volcanic eruptions, and man-made sulphur emissions over a decade as compared with the results using the four policy scenarios which give, over the next decade, changes ranging from  $0.41$  to  $0.56\text{ Wm}^{-2}$ . For solar flux variations, it is assumed that the variability over a decade, when averaged over the eleven-year solar cycle, should be less than the longer-term change. The earlier discussion indicates that over a decade the solar flux variability could modify the radiative forcing by  $\pm 0.1\text{ Wm}^{-2}$  and one large volcanic eruption in a decade could cause a decrease of  $0.2\text{ Wm}^{-2}$ . The global-mean effect of sulphur emissions on cloud albedos was estimated to be up to  $0.15\text{ Wm}^{-2}$  per decade, but, on a decadal scale, not even the sign of the effect is certain. Since both the volcanic and sulphate effects do not act globally, the possible compensations between increased greenhouse forcing and possible decreases from the other effects may be even greater regionally, whilst in other regions, such as in the southern hemisphere, the impact of sulphur emissions may be very small.



**Figure 2.6:** Comparison of different radiative forcing mechanisms for (a) a 10 year period, and (b) a 50 year period in the future. The greenhouse gas forcings are for the periods 1990–2000 and 2000–2050 respectively using the four policy scenarios. Forcings due to changes in solar radiation and sulphur emissions could be either positive or negative over the two periods.

While other effects could greatly amplify or negate the greenhouse-gas-induced radiative forcing over any given decade, the effects of such forcings over a longer time period should generally be much smaller than the forcing expected from the greenhouse gases. This is shown in Figure 2.6(b) for the changes in radiative forcing over a 50-year period. The four policy scenarios lead to changes in forcing of between  $1.3$  and  $3.5\text{ Wm}^{-2}$  for the period 2000–2050. The effects from solar variability, volcanic

eruptions and man-made sulphur emissions are likely to be much smaller. The prior discussion suggests a change in radiative forcing of  $0.2 \text{ Wm}^{-2}$  from solar variability could occur over several decades. In the unlikely case of one major volcanic eruption per decade, a resulting net decrease in radiative forcing of  $0.2 \text{ Wm}^{-2}$  could be sustained over a 50 year period. The effect of man-made sulphur emissions is again highly uncertain but using the earlier estimates it could be up to  $0.75 \text{ Wm}^{-2}$  of either sign. Effects on radiative forcing from changes in surface characteristics should be less than  $0.1 \text{ Wm}^{-2}$  over this time period.

In addition to the effects from other forcings that oppose or reinforce the greenhouse gas forcing, there are also decadal-scale climate changes that can occur without any changes in the radiative forcing. Non-linear interactions in the Earth-ocean-atmosphere system can result in unforced internal climatic variability (see e.g. Section 6.5.2). As a result of the combined effects of forced and unforced effects on climate a range of unpredictable variations of either sign will be superimposed on a trend of rising temperature.

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