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The effect of human activity on radiative forcing of climate change: a review of recent developments

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Abstract

Human activity has perturbed the Earth's energy balance by altering the properties of the atmosphere and the surface. This perturbation is of a size that would be expected to lead to significant changes in climate. In recent years, an increasing number of possible human-related climate change mechanisms have begun to be quantified. This paper reviews developments in radiative forcing that have occurred since the second assessment report of the Intergovernmental Panel on Climate Change (IPCC), and proposes modifications to the values of global-mean radiative forcings since pre-industrial times given by IPCC. The forcing mechanisms which are considered here include those due to changes in concentrations of well-mixed greenhouse gases, tropospheric and stratospheric ozone, aerosols composed of sulphate, soot, organics and mineral dust (including their direct and indirect effects), and surface albedo. For many of these mechanisms, the size, spatial pattern and, for some, even the sign of their effect remain uncertain. Studies which have attributed observed climate change to human activity have considered only a subset of these mechanisms; their conclusions may not prove to be robust when a broader set is included.

Keywords: greenhouse gases; climate forcing; aerosols

1. Introduction

There has been increasing attention given to the impact of human activity on climate. The field has developed rapidly, as can be illustrated by reference to some past reviews. In the early 1970s, attention was predominantly focused on the effects of increased atmospheric concentrations of carbon dioxide (Matthews et al., 1971; SMIC, 1971). It was also known that changes in the atmospheric concentrations of sub-micron particles ('aerosols') could have an impact on climate directly, by interacting with solar radiation (wavelengths 0.2 to 4 μm), and indirectly, by altering the properties of clouds; however, the size and sign of these effects were matters of debate.

Later in the 1970s, it was recognised that the atmospheric concentrations of other gases, such as methane and the chlorofluorocarbons (CFCs), had increased markedly. Like carbon dioxide, they can absorb and emit thermal infrared radiation (wavelengths 4 to 100 μm), and enhance the Earth's natural greenhouse effect. The cumulative impact of these 'new' gases was more confidently quantified during the 1980s; they were believed to enhance the warming effect due to carbon dioxide alone by about 50% (e.g., Clark, 1982; Dickinson and Cicerone, 1986; Ramanathan et al., 1987). Much less attention was focused on aerosols in this period, but the

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potential impact of changes in tropospheric and stratospheric ozone began to be recognised (e.g., Clark, 1982; Dickinson and Cicerone, 1986; Ramanathan et al., 1987).

In the late 1980s, there was a resurgence of interest in aerosols (see e.g., Andreae, 1995; Charlson, 1997). This was driven partly by an improved understanding of the aerosol sources and characteristics and partly by the development of three-dimensional numerical models (so-called chemical transport models (CTMs)) which generate estimates of global aerosol distributions. The first scientific assessment of the Intergovernmental Panel on Climate Change (IPCC) (IPCC, 1990) reported the possibility of a significant aerosol effect but remained tentative about its importance; aerosols occupied an increasingly prominent role in subsequent IPCC reports (IPCC, 1992, 1994, 1995). The initial focus on sulphate aerosols broadened to cover other aerosol types, such as soot, organics and dust. Interest in the role of ozone returned in the 1990s, following the unambiguous detection of global-scale stratospheric ozone depletion (e.g., WMO, 1995), and the use of CTMs to model the distribution of tropospheric ozone and its changes since pre-industrial times.

Potential climate change mechanisms are characterised here using the change they cause in the planetary energy balance (this is known as the 'radiative forcing', in W m^{-2}) if all other tropospheric and surface variables are held fixed (see Appendix A). Although the effects of human activity are the focus here, they act against a backdrop of natural climate change mechanisms, such as changes in solar output and volcanic eruptions (e.g., Hansen and Lacis, 1990; IPCC, 1994, 1995; Rowntree, 1998). Further, the earth-atmosphere climate system is a non-linear system; climate changes need not have an external forcing agent, nor need the response to those forcings be linear (e.g., Palmer, 1993; Hansen et al., 1997c). Additionally, climate change can occur due to non-radiative processes, such as land use changes which alter the roughness of the surface or its ability to evaporate water (e.g., Henderson-Sellers, 1995); such changes are not considered here.

To calculate the radiative forcing due to a particular species, it is necessary to know (i) its three-dimensional distribution and how this has changed over a given period and (ii) how it interacts with solar and thermal infrared radiation. For some constituents, this is relatively straightforward. Carbon dioxide is long-lived [with an adjustment time of order decades (IPCC, 1995)] compared to the timescales with which the atmospheric circulation redistributes molecules. Its atmospheric distribution is therefore quite homogeneous and only a few measurements are required across the globe to characterise its growth rate. Pre-industrial concentrations can be inferred from measurements of air trapped in ice cores (IPCC, 1994). Its radiative properties are well-characterised and so its interaction with solar and thermal infrared radiation can be calculated with relatively high confidence. The same is true for all the so-called 'well-mixed' greenhouse gases, which, in addition to carbon dioxide, include methane, nitrous oxide and the CFCs.

Problems occur for constituents such as ozone and aerosols. In the lower troposphere, these are short-lived (of order days to weeks (IPCC, 1995)); because their emission sources (or those of their gaseous pre-cursors) are localised in, for example, industrialised regions or biomass-burning regions, the distributions are vertically and horizontally inhomogeneous. Thus, a high density of observations would be required to define the distribution. Such observations do not exist; even if they did, there is no reliable method to allow anything but a crude estimate of their pre-industrial concentration. In the case of aerosols, not only must the distribution be known, but knowledge of their composition is required to determine the sign of their climatic effect (e.g., Charlson and Heintzenberg, 1995). Similarly, the sign of the climate effect of stratospheric ozone change depends on details of the vertical distribution of this change (e.g., Ramanathan and Dickinson, 1979).

Because of limited present-day observations and the scarcity of pre-industrial data, CTMs have been increasingly used in climate studies over the past decade (e.g., Langner and Rodhe, 1991; Chuang et al., 1997; Feichter et al., 1997; Lelieveld et al., 1997). These models require the spatial, time-varying, distribution of emissions of all relevant species and the atmospheric winds, which move the species; they must include the chemical reactions, which interconnect species, and cloud and rain processes, which determine both the rate of some chemical

reactions and remove species from the atmosphere. Much progress has been made, but questions remain on, for example, the size of emissions sources (e.g., Lelieveld et al., 1997), and on how the vertical transport is represented; much of this transport is achieved by convection which occurs on much smaller scales (of order kilometres) than the typical model grid spacing (of order 100 km).

An additional complication is that the concept of radiative forcing assumes a clear distinction between radiative forcing mechanisms and climate feedback mechanisms but a clear separation is not always possible; for example, if part of the stratospheric ozone change were to be due to temperature and circulation changes resulting from some other forcing (such as increases in carbon dioxide), it could be argued that the radiative impacts of the ozone changes are a feedback, in much the same way that changes in tropospheric water vapour are conventionally viewed (see Appendix A).

This paper will review developments in radiative forcing since the IPCC second scientific assessment (IPCC, 1995). For each climate change mechanism, the global-mean radiative forcing since the mid-19th century ('pre-industrial') is presented as a central value with a range based, where possible, on the spread of published values; this range is not intended to be a strict error bar and each forcing is accompanied by a subjective 'confidence level' which indicates our confidence that the true forcing lies within the specified range, given the difficulties in characterising that forcing. Examples of calculations of the geographical distribution of forcing are also shown.

2. Climate change mechanisms

2.1. Well-mixed greenhouse gases

The radiative forcing due to increased concentrations of well-mixed greenhouse gases is the best characterised forcing. Recent estimates for the forcing since pre-industrial times lie within 7% of 2.4 W m^{-2}, which is due to carbon dioxide (about 60%), methane (about 20%), the CFCs and related species (about 12%) and nitrous oxide (about 6%) (IPCC, 1995; Bengtsson, 1997; Hansen et al., 1997b; Lelieveld et al., 1998; Myhre et al., 1998b). Some of the 7% spread results from differences in radiative transfer calculations, but some also results from different specifications of the pre-industrial concentrations of species. The uncertainty for individual molecules can be higher; for example, the CFC-11 forcing may be more than 10% higher than the IPCC (1995) estimate (Christidis et al., 1997; Myhre and Stordal, 1997; Hansen et al., 1997a), and there is a divergence of 20% in estimates of the methane forcing (IPCC, 1994; Myhre et al., 1998b). The forcing appears to be robust to assumptions about spatial and temporal averaging (Myhre and Stordal, 1997) and uncertainties in the spectroscopy of carbon dioxide, methane and nitrous oxide (Pinnock and Shine, 1998), but uncertainty in the specification of the tropopause position may add a more significant uncertainty ($\sim$ 10%) (Myhre and Stordal, 1997; Freckleton et al., 1998). IPCC (1995) estimated an uncertainty of $\pm$10% which we retain. Fig. 1a shows a calculated distribution of radiative forcing due to the well-mixed greenhouse gases. The pattern is determined mainly by the variations in cloudiness (particularly high-level cloud) and surface temperature.

2.2. Stratospheric ozone

The forcing due to halocarbon-driven stratospheric ozone depletion (e.g., WMO, 1995, 1999) has generally been treated separately from the effects of tropospheric ozone increases. This is mainly because the stratospheric changes have largely occurred in the last two decades and better (but still not fully adequate) observational data exist; the tropospheric ozone changes have occurred over about a century and analysis has been largely based on CTMs.

Calculation of the stratospheric ozone radiative forcing is not straightforward. Ozone depletion causes an increase in solar radiation reaching the troposphere—a positive forcing. The decreased emission from ozone's thermal infrared bands causes a negative forcing. The loss of ozone also cools the stratosphere (the 'stratospheric adjustment' process described in Appendix A); the consequent decrease in thermal infrared emission is a further negative forcing. The balance between these processes, and hence the sign of the forcing, is sensitive to the height profile of the ozone loss (e.g., Ramanathan and Dickinson, 1979; Forster and Shine, 1997; Hansen et

al., 1997a) and on the way stratospheric temperature changes are calculated (Zhong et al., 1996; Shine et al., 1998). In general, ozone loss in the lower stratosphere gives a negative radiative forcing and ozone loss in the upper stratosphere gives a positive forcing.

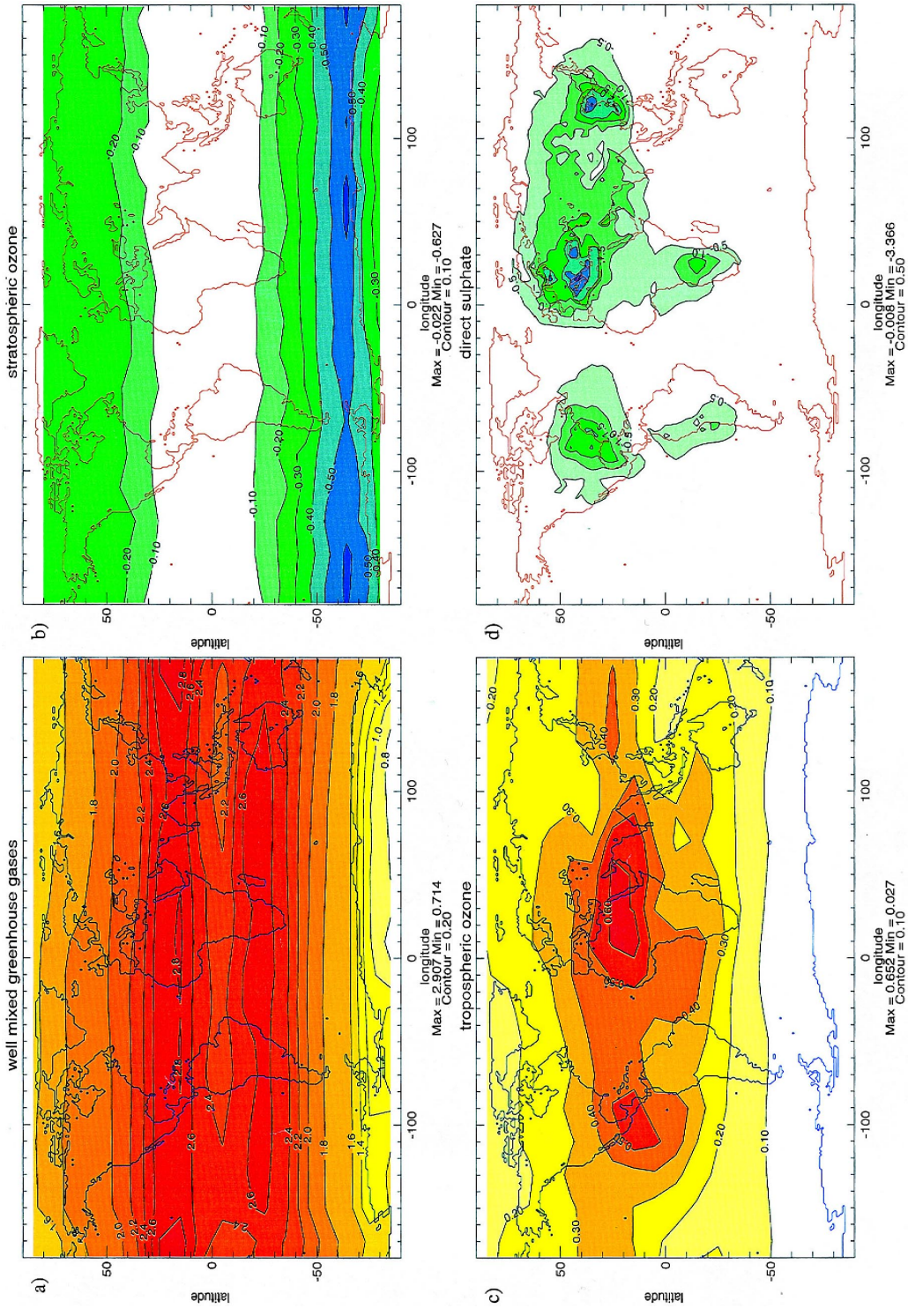
Since IPCC (1995), there have been several studies using a range of techniques to derive the stratospheric ozone radiative forcing. The first estimate using a general circulation model (GCM) (albeit with quite low vertical resolution), with alternative sets of satellite-derived ozone changes, has produced the most negative forcing of recent studies, in the range from -0.13 to -0.19 W m^{-2} decade^{-1} since the late 1970s (Hansen et al., 1997a). Models using observed ozone changes but with simpler methods to derive stratospheric temperature change (Forster and Shine, 1997; MacKay et al., 1997) have obtained forcings between -0.05 and -0.13 W m^{-2} decade^{-1}. An attempt to derive the forcing using satellite-observed stratospheric temperature changes as well as observed ozone changes (Zhong et al., 1996), obtained a forcing of -0.02 W m^{-2} decade^{-1}; making such an estimate is difficult because of the poor vertical resolution of satellite observations and a similar calculation but using radiosonde-based temperature trends yielded a forcing of -0.1 W m^{-2} decade^{-1} (Shine et al., 1998). Some of the divergence in the estimates is due to differences in the details of the radiative forcing calculations but most is probably due to the difficulties in defining the vertical and latitudinal profiles of ozone loss and in representing the consequent stratospheric temperature change.

One study (Myhre et al., 1998a) derived ozone changes from a chemical model, rather than from observations; because of a larger relative loss in the upper stratosphere, a positive forcing of 0.02 W m^{-2} decade^{-1} for the 1980s was obtained. Given the difficulties in modelling the ozone change in the lower stratosphere, particularly at mid-latitudes (WMO, 1995), it is not clear how much weight should yet be given to this estimate; nevertheless, it sends a clear warning that there are significant uncertainties in deriving the climatic effect of ozone loss.

An additional uncertainty is that the consequent increased penetration of UV radiation to the troposphere is expected to increase the production rate of OH which may have a subsequent impact on radiatively active species (e.g., Bekki et al., 1994; Toumi et al., 1994; WMO, 1999). The radiative forcing consequences have not yet been assessed in recent 3D CTMs but they may already have been taken partially into account in calculations to the extent that observed methane concentrations have responded to the UV change.

IPCC (1995) gave the stratospheric ozone forcing as -0.1 W m^{-2} with a factor of two uncertainty. More recent studies have included the ozone loss up to the mid-1990s (Forster and Shine, 1997; Hansen et al., 1997a). Forster and Shine (1997) extended the calculation back to 1964 using ozone changes deduced from surface-based observations and assumed that the decadal rate of change of forcing from 1979

Fig. 1. Spatial patterns of estimates of radiative forcing due to a number of mechanisms resulting from human activity: (a) well-mixed greenhouse gases; (b) stratospheric ozone; (c) tropospheric ozone; (d) direct effects of sulphate aerosols; (e) direct effects of fossil-fuel soot aerosols; (f) mineral dust from disturbed soils; (g) the indirect effects of sulphate aerosols. (a) used the pre-industrial to present day changes in concentrations of carbon dioxide, methane, nitrous oxide, CFC-11 and CFC-12 given in IPCC (1995), and the University of Reading radiation schemes and climatology (Berntsen et al., 1997); (b) used the zonal-mean stratospheric ozone depletions from 1979–1994 given in WMO (1995) with the 'SBUV' ozone depletion profile as described by Forster and Shine (1997); (c) used the pre-industrial to present day changes in ozone simulated by the University of Oslo chemical-transport model and the University of Reading radiation schemes (Berntsen et al., 1997); (d) used the Langner and Rodhe (1991) 'slow-oxidation' case for the pre-industrial to present-day sulphate aerosol changes and the UK Meteorological Office Unified Model (Haywood et al., 1997b); (e) is a speculative forcing pattern for fossil-fuel soot which assumed that the mass of soot is a fixed fraction (0.075) of the sulphate distribution described in (d) and used the UK Meteorological Office Unified Model (Haywood et al., 1997b); (f) is the forcing from the modelled mineral-dust raised from soils disturbed by human activity and used the NASA Goddard Institute of Space Studies GCM (Tegen et al., 1996)—the strong forcing over the Himalayas may be biased high because the model surface albedo is too high in this region and the dust transport to this region is believed to be too high (Ina Tegen, personal communication, 1997); (g) is one example of a modelled pattern of indirect sulphate forcing derived using the same sulphate distribution as in (d) and different functions linking sulphate aerosol mass and cloud droplet size over continents and oceans and used the UK Meteorological Office Unified Model (Jones and Slingo, 1997).


a)
well mixed greenhouse gases
Max = 2.907 Min = 0.714
Contour = 0.20
b)
stratospheric ozone
Max = -0.022 Min = -0.627
Contour = 0.10
c)
tropospheric ozone
Max = 0.652 Min = 0.027
Contour = 0.10
d)
direct sulphate
Max = -0.008 Min = -3.366
Contour = 0.50
latitude
longitude

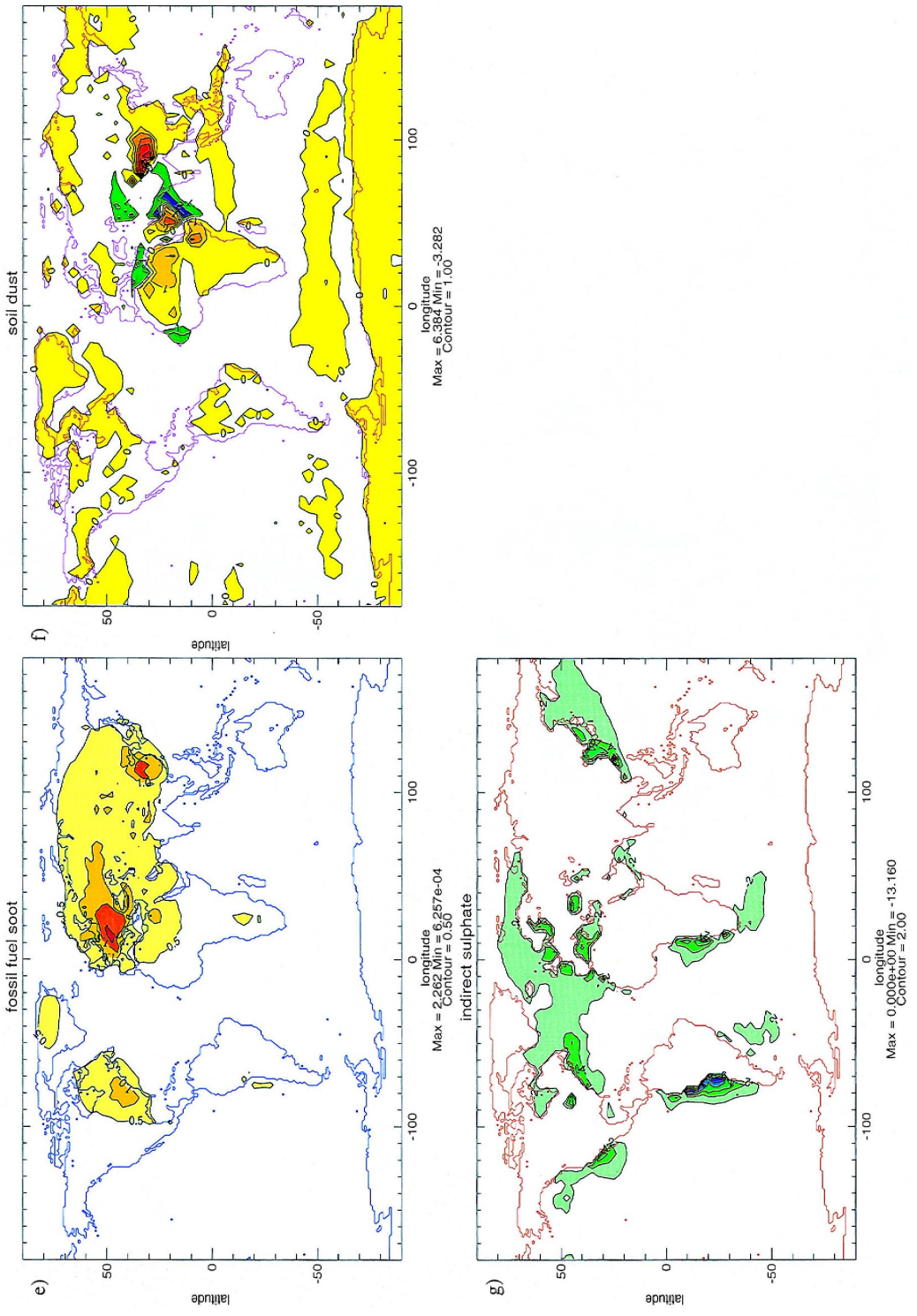


Fig. 1 (*continued*).

to 1991 was sustained to the mid-1990s; they obtained a global-mean forcing due to stratospheric ozone loss of about -0.23 W m^{-2}.

More recently, WMO (1999) have presented updates to the ozone trends analyses from 1979 to 1997. We have used these trends together with the simple method for estimating forcing given in Forster and Shine (1997) to derive a revised value. For the period 1979–1997, we obtain a value of -0.15 W m^{-2}; we now choose not to include the pre-1979 depletion, as the serious lack of knowledge of the heights of this earlier change mean that even the sign is uncertain. The spread in forcings, from other studies using observed ozone changes, implies a range of about ± 0.12 W m^{-2}. Both the size and the uncertainty of the forcing are higher than given by IPCC (IPCC, 1995) and we retain their 'low' confidence level. Fig. 1b shows an estimate of the geographical distribution of the stratospheric ozone radiative forcing. The forcing becomes more negative towards the poles, because ozone depletion is generally greatest there; the pattern enhances the latitudinal gradient due to the well-mixed greenhouse gases (Fig. 1a).

2.3. Tropospheric ozone

During the late 19th century a few, mostly European, observations of the surface-level ozone were made; these indicate that large changes in surface ozone have occurred since then—over Europe concentrations have more than doubled (e.g., Marenco et al., 1994). These increases are due to emissions of oxides of nitrogen, carbon monoxide and hydrocarbons, much of it from fossil fuel and biomass burning (e.g., IPCC, 1995). Recent estimates of radiative forcing have been mostly based on output from CTMs (Hauglustaine et al., 1994; Chalita et al., 1996; Forster et al., 1996; Berntsen et al., 1997; Roelofs et al., 1997; Van Dorland et al., 1997; Brasseur et al., 1998b, Haywood et al., 1998; Stevenson et al., 1998) and also from present and pre-industrial observations (Marenco et al., 1994); satellite-inferred ozone increases due to biomass burning have also been used (Portmann et al., 1997).

Comparison of estimates is not straightforward because of methodological differences (for example, whether clouds are included in the calculation). When these differences are accounted for (Berntsen et al., 1997), the global-mean tropospheric ozone radiative forcing from recent studies are in the range 0.35 ± 0.15 W m^{-2}; the range seems mostly due to differences in the CTM-calculated ozone change. The forcing and range are slightly smaller than given by IPCC (1995). Most of the recent estimates lie close to or in the lower half of this range; however, until the causes of the divergence in the simulated ozone change are identified, it would be premature to revise the range further. Indeed, even where they agree in the global-mean net forcing the available results differ quite significantly in the impact of clouds on the radiative forcing. For example, Haywood et al. (1998) find the shortwave forcing triples on including clouds, whereas Berntsen et al. (1997) and Van Dorland et al. (1997) report only a 40% enhancement; this difference may be due to differences in the vertical and horizontal distribution of ozone change or the way clouds are included in the calculations. IPCC (1995) gave a 'low' confidence to the forcing estimate, which we retain. Only Stevenson et al. (1998) have so far reported the time evolution of the radiative forcing since pre-industrial times, and find the forcing to be approximately linear in NO_x emissions.

The geographical distribution of this forcing (Fig. 1c) is much more inhomogeneous than that of the well-mixed greenhouse gases (Fig. 1a). The annual mean forcing reaches a peak over north Africa due to a coincidence of relatively large ozone changes with the region where those changes are most effective due to cloudless skies, a warm surface and high surface reflectance (Berntsen et al., 1997).

2.4. Direct effects of aerosols

The atmospheric aerosol burden has increased due to human activity. The burning of fossil fuel leads to emission of sulphur dioxide which, on a timescale of days, oxidises to gaseous sulphuric acid which then forms aerosols (e.g., Charlson and Heintzenberg, 1995; IPCC, 1995; Charlson, 1997). Small amounts of elemental, or black, carbon (referred to here as 'soot') and organic carbon are also emitted (e.g., Cooke and Wilson, 1996; Liousse et al., 1996); biomass burning leads to emissions of both soot and

organic aerosol precursors (Penner et al., 1992; Hobbs et al., 1997), and land use changes has led to an increase in airborne mineral dust (Tegen et al., 1996).

Sulphate aerosol distributions have been derived from CTMs (e.g., Langner and Rodhe, 1991; Chuang et al., 1997; Feichter et al., 1997; Lelieveld et al., 1997). There is a lack of data to evaluate these distributions comprehensively, but at least at the surface, there is some tendency to underestimate the sulphate concentration in polluted regions, possibly because oxidation on other aerosol components is neglected (Lelieveld et al., 1997).

Most of the recent estimates of the direct forcing of sulphate aerosols (Boucher and Anderson, 1995; Chuang et al., 1997; Feichter et al., 1997; Graf et al., 1997; Haywood et al., 1997b; Van Dorland et al., 1997; Myhre et al., 1998c) fall in the range from about -0.25 to -0.4 W m^{-2}. However, some are much more negative. Kiehl and Rodhe (1995) obtained a forcing of -0.66 W m^{-2}, partly because higher sulphur emissions were assumed; Haywood and Ramaswamy (1998) found a forcing of -0.8 W m^{-2} not because of a high sulphate burden, but because relatively more was at lower altitudes, and hence at higher relative humidities. Hygroscopic aerosols, such as sulphates, grow as the relative humidity is increased, to sizes (between 0.1 and 1 μm) which are better able to interact with solar radiation (IPCC, 1995). Relative humidity generally decreases with altitude; hence, for a given sulphate mass, the radiative forcing becomes less negative if the aerosol height is increased (Haywood and Shine, 1997). Penner et al. (1998) also obtained -0.8 W m^{-2} and also highlighted the sensitivity of the forcing to the representation of relative humidity in their model.

A related issue is that most current CTMs typically have horizontal resolutions of about 250 km. Relative humidity varies on much smaller scales than this and, due to the nature of the dependence of aerosol scattering on relative humidity, neglect of this variation leads to a systematic underestimate of the magnitude of the forcing (Haywood et al., 1997a). The global-scale importance of this has only been crudely estimated (Haywood and Ramaswamy, 1998), but it may make the forcing around 10% more negative.

A common assumption in simple radiation models of the aerosol effect is that sulphate aerosol in cloudy regions contributes negligibly to the radiative forcing, as it adds only a small increment to the total optical depth. However, results from more sophisticated radiation schemes are not in agreement. Using the same technique for computing the forcing from cloudy skies, Haywood et al. (1997b), Haywood and Ramaswamy (1998) and Myhre et al. (1998c) find that cloudy regions contribute 5, 11 and 27%, respectively. Boucher and Anderson (1995) and Van Dorland et al. (1997) also obtain values in excess of 20%, although Haywood and Shine (1997) have shown that their method of deriving the cloudy sky forcing may not be reliable. Liao and Seinfield (1998), using a single column model, have explored the conditions under which sulphate aerosols contribute to forcing in cloudy skies; the reduction in forcing, as cloud thickness is increased, is not as rapid as that found in similar calculations by Haywood and Shine (1997). The spread in these results may relate to the representation of clouds in these models, or to details of the radiative transfer models; the cause of the difference needs to be resolved. Certainly, the spread in the zenith angle dependence of the forcing reported in the Boucher et al. (1998) intercomparison may account for some of these differences.

IPCC (1995) gave a value of radiative forcing of -0.4 W m^{-2}, with a factor of 2 uncertainty. The range of recent results is no smaller than this, so we retain the IPCC estimate and the 'low' confidence level. Fig. 1d shows one estimate of the geographical distribution of sulphate forcing. Because of the short (several-day) lifetime of the sulphate aerosol, the effects are most marked in regions immediately downwind of major industrial centres.

Van Dorland et al. (1997) have also reported the first tentative estimate of a forcing from nitrate aerosols; in the northern hemisphere, the nitrate forcing is found to be an order of magnitude lower than the sulphate forcing.

Soot is also of potential importance; because it is an effective absorber of solar radiation, it reduces the planetary reflectance causing a positive forcing. There have been several recent studies on the role of soot (Chylek and Wong, 1995; Chylek et al., 1995; Schult et al., 1997; Haywood et al., 1997b; Haywood

and Ramaswamy, 1998; Penner et al., 1998; Myhre et al., 1998c) but results are still preliminary.

There are a number of problems: the two available CTM studies (Cooke and Wilson, 1996; Liousse et al., 1996) disagree by a factor of 2 on the soot burden; it is difficult to characterise the sources of soot emission and measure the amount of absorption, particularly for the sub-micron particles that are of most climatic importance (IPCC, 1995; Heintzenberg et al., 1997; Bond et al., 1998); and the radiative forcing depends strongly on the assumed aerosol size (Chylek and Wong, 1995; Haywood and Ramaswamy, 1998). Soot is most effective over reflecting surfaces (such as deserts, snow and clouds) rather than dark surfaces (such as the oceans) as it has a greater influence on the planetary reflectance. Also, the higher the soot altitude, the stronger its effect is, as it is more likely to be above cloud (Haywood et al., 1997b; Haywood and Ramaswamy, 1998). The state of mixing of the soot with other aerosols and cloud droplets is also important (Chylek et al., 1996; Heintzenberg and Wendisch, 1996; Haywood et al., 1997b). If soot exists separately (an 'external mixture'), it is less effective than if it is incorporated within other aerosols or cloud droplets (an 'internal mixture'). There are only sparse data, mostly at the surface, with which to constrain models (e.g., Heintzenberg et al., 1997; Myhre et al., 1998c).

An added complication is that, in mass terms, there are two approximately equal sources of soot: fossil fuel burning and biomass burning (Cooke and Wilson, 1996; Liousse et al., 1996). Some studies include the soot from one or other of these sources, while others include both. To avoid confusion, we first consider the fossil-fuel soot forcing which is estimated to be about $+0.2$ W m^{-2} with an uncertainty of at least a factor of 3 (Haywood et al., 1997b; Haywood and Ramaswamy, 1998; Penner et al., 1998; Myhre et al., 1998c). This estimate is twice that given by IPCC (1995) because, at that time, the only study available considered just the clear sky component. The IPCC (1995) confidence level of 'very low' is retained. A significant point here is that estimates of the global-mean forcing due to fossil-fuel soot and sulphate overlap in size but are opposite in sign. Fig. 1e shows a crude estimate of the geographical pattern of radiative forcing from fossil-fuel soot by Haywood et al. (1997b) who used the same geographical distribution as sulphate aerosols (as used in Fig. 1d) and applied a simple scaling to derive a soot mass; more recent studies (Haywood and Ramaswamy 1998; Penner et al., 1998) have used CTM-derived soot distributions. Although the soot and sulphate forcing patterns show some similarities, there are significant differences because soot is less effective over oceans but more effective over continents.

The forcing due to biomass-burning aerosols (which are largely composed of organic carbon and soot) has been computed in some studies by considering the net effect of all aerosol types resulting from the burning (Penner et al., 1992, 1998; Chylek and Wong, 1995; Hobbs et al., 1997); other studies have looked specifically at the soot component (Schult et al., 1997; Haywood and Ramaswamy, 1998). Hobbs et al. (1997) used observations during the Amazonian biomass-burning season to characterize the optical properties; their estimate of the global-mean forcing was -0.3 W m^{-2}. Penner et al. (1998) have produced the first estimate based on CTM-derived aerosol distributions; depending on the choice of aerosol size distribution, and whether the aerosols were present in cloudy regions, their global-mean forcing ranged from -0.14 to -0.23 W m^{-2}.

Satellite observations of biomass burning plumes confirm that the aerosols enhance reflectance (Christopher et al., 1996) although, interestingly, they also indicate a non-negligible effect on the thermal infrared; this causes a positive forcing which offsets part of the shortwave effect. Christopher et al. (1998) have extended this work to further quantify the negative radiative forcing associated with smoke close to the burning regions and discuss the ways newer satellite data may be used to deduce forcing averaged over a day and over broader regions. The interpretation of satellite data and in situ observations plays a very important role in increasing our understanding of radiative forcing, but it is important that it is made clear what the radiative forcing is relative to. If the forcing is derived for a smoke plume relative to surrounding regions, then this background region may itself have enhanced aerosol loading. If the derived forcing is relative to a hypothetical non-aerosol state, then this value includes the contribution of natural aerosols. A difficult challenge will be to use satellite data to separate out the

anthropogenic aerosol forcing from the total aerosol forcing; this will probably only be achievable in conjunction with both chemical-transport and radiation models and in situ observations.

Since some of the biomass burning predates the early 19th century (Andreae, 1995), it has been estimated that the forcing should be reduced by between one-third and one-half to give the change since 1850. IPCC (1995) gave a global-mean forcing due to the direct effect of biomass burning since the 19th century as -0.2 W m^{-2} with a factor of 3 uncertainty. On available evidence, it would be premature to revise this or the 'very low' confidence level.

The final aerosol category considered here is mineral dust. Mineral dust blown from arid and semi-arid regions is a major contributor to the total natural atmospheric aerosol burden (Andreae, 1995; IPCC, 1995). Changes in land use are estimated to account for between one-quarter and one-half of the mineral dust burden (Sokolik and Toon, 1996; Tegen et al., 1996). Forcing calculations are as yet in their infancy. Mineral dust absorbs as well as scatters solar radiation and, over surfaces with sufficiently high reflectance, it can cause a positive forcing; the degree of the mineral dust absorption is not well characterised (Sokolik and Toon, 1996). Some of the dust is sufficiently large for the aerosols to have a significant impact on thermal infrared radiation and act rather like greenhouse gases, causing a positive forcing.

On a global average, the only available estimate (Tegen et al., 1996) shows a near cancellation of the solar forcing (-0.25 W m^{-2}) and the thermal infrared forcing ($+0.34$ W m^{-2}), giving a net forcing of $+0.09$ W m^{-2}; locally, however (see Fig. 1f), the forcing is much higher, varying from over $+5$ W m^{-2} to about -3 W m^{-2} between Arabia and the Arabian Sea. Large (natural) dust outbreaks change the aerosol burden sufficiently for it to be clearly detected in solar and thermal infrared satellite observations (Ackerman and Chung, 1992); these observations confirm that dust causes a positive forcing over desert and a negative forcing over ocean, and help give some qualitative confidence in the models.

As only one estimate is available, and since the net forcing is a small residual of two larger terms, it is difficult to estimate an uncertainty. Sokolik and Toon (1996) estimate an uncertainty in the visible optical depth of about a factor of 3 and Miller and Tegen (1998) show that a 10% variation in the single scattering albedo of the mineral dust at solar wavelengths can alter the sign of the net forcing. We take the Tegen et al. (1996) value of $+0.1$ W m^{-2} and assign a rather arbitrary uncertainty of ± 0.4 W m^{-2}, with a 'very low' confidence. Much work is required to improve the accuracy of the forcing, even to the extent of needing to confirm the sign. Nevertheless, in terms of direct forcing, the size of the mineral dust forcing may be similar to the other aerosol components considered here.

Most studies have considered aerosol components in isolation but there is a growing realisation that actual aerosol populations are a complex mix of different components, some natural and some anthropogenic in origin (e.g., Hegg et al., 1997; Tegen et al., 1997; Murphy et al., 1998). Further, the net effect of changes in all aerosol is unlikely to be equal to the sum of the individual components. The net effect depends on the degree to which the aerosols are internally or externally mixed as discussed earlier. It has also been proposed (Dentener et al., 1996) that the presence of mineral dust may markedly affect the rate at which sulphate and nitrate aerosols form; this will affect both the total aerosol burden and the radiative effect since sulphate forming on large mineral dust particles is likely to be less radiatively effective than a sub-micron sulphate aerosol on its own. Finally, West et al. (1998) have shown that for internally-mixed aerosols, the radiative forcing may not even increase monotonically with sulphate loading, because of interactions between the sulphate and other components of the aerosol population, such as ammonium, nitrate and water.

2.5. Indirect aerosol effects via changing cloud properties

Cloud droplets form on to cloud condensation nuclei (CCN) which are a subset of the total aerosol population. There is abundant evidence that cloud properties are affected by the source (and hence aerosol content) of the air in which it forms (e.g., Taylor and McHaffie, 1994; Garrett and Hobbs, 1995; Hudson and Li, 1995) and it has long been proposed that human-induced changes in CCN could

alter cloud properties (e.g., SMIC, 1971; Twomey et al., 1984). Increases in CCN numbers will, all else being equal, lead to a larger number of smaller droplets which will increase the cloud's solar reflectance, giving a negative radiative forcing. The decreased droplet radius might also decrease the cloud's ability to precipitate leading to an increased cloud lifetime and liquid water content (Pincus and Baker, 1994; Boucher et al., 1995; Lohmann and Feichter, 1997) causing an additional negative forcing; since the removal rate of aerosols also depends on these cloud processes, the aerosol burden might itself be influenced (Lohmann and Feichter, 1997). We refer to all forcings resulting from changed cloud properties due to aerosol-induced effects on droplet radius as the indirect aerosol forcing.

There are now several studies employing different techniques which have investigated the indirect forcing due to sulphate aerosol from changes in droplet size alone. Global-scale models have been used with CTM-derived sulphate climatologies (Chuang et al., 1997; Feichter et al., 1997; Jones and Slingo, 1997); very-high resolution models (with grid spacings of about 50 m), which have a much more detailed representation of cloud microphysics than the global models, have also been used in association with observed cloud amounts (Kogan et al., 1996, 1997). Satellite-derived inter-hemispheric differences in droplet radii in low clouds (Boucher, 1995) have also been used, although this study does not strictly separate out the sulphate effect nor the human-induced component. For all these studies, the global-mean indirect sulphate forcing lies between about -0.5 and -1.5 W m^{-2}. There may be a factor of 2 amplification of this forcing due to the additional impact on cloud amount and liquid water content (Pincus and Baker, 1994; Boucher et al., 1995; Lohmann and Feichter, 1997). Lohmann and Feichter (1997) find that the indirect sulphate forcing may be as negative as -4.8 W m^{-2}, although alternative representations of cloud cover in their model weakens the effect to -1.4 W m^{-2}.

There are many difficulties in quantifying the indirect forcing (e.g., IPCC, 1994, 1995; Schwartz and Slingo, 1996; Pan et al., 1998). The aerosol mass, derived from CTMs, has to be related to the cloud droplet concentration in some way. The nature of this relationship is not clear and is likely to differ between continental areas, with high background aerosol concentrations, and remote marine locations (e.g., Jones and Slingo, 1997; Chuang et al., 1997; Feichter et al., 1997). Most studies have concentrated on the role of sulphates but other aerosol components may be important. Additional aerosols are most effective when the pre-existing aerosol concentrations are low but the effect saturates when aerosol concentrations are high (Twomey et al., 1984; Hegg, 1994; Schwartz and Slingo, 1996; Kogan et al., 1997). It matters whether the additional sulphate is forming fresh aerosol or condensing on existing particles (Chuang et al., 1997) and what the background aerosol loading is (Graf et al., 1997). If the 'background' aerosol, which acts to reduce the sulphate effect, is in itself substantially due to human activity, then while the additional sulphate itself may appear to contribute a lesser indirect forcing, this would be because the other aerosols had already caused a substantial indirect effect (Chuang et al., 1997; Jones and Slingo, 1997).

A role for other aerosol types in the indirect forcing has been proposed, such as biomass-burning aerosols (Penner et al., 1992), other organic aerosols (Novakov and Penner, 1993; Hegg, 1994) and nitrate aerosols (Kulmala et al., 1995); mineral dust aerosols may become more active as CCN when coated by sulphate (Dentener et al., 1996). Quantification of the effect of these other aerosols is either very preliminary or has not been made. As with the aerosol's direct effect, the effect of the total change in aerosol population may differ significantly from the sum of the individual components.

Because of the many difficulties, IPCC (1995) refrained from giving a central value to the indirect forcing, indicating that it lies in the range 0 to -1.5 W m^{-2}, with 'very low' confidence. While we retain the 'very low' confidence, recent studies have explored a large parameter space of uncertainty and all indicate a negative forcing and, when all aerosol types are considered, one that may be more negative than -1.5 W m^{-2}; we adopt a value for the indirect forcing due to all aerosols of -1 W m^{-2} with at least a factor of two uncertainty. The geographical distribution of this forcing is also uncertain. Fig. 1g shows one calculation (from Jones and Slingo, 1997) of the pattern of forcing which uses the same sulphate distribution as was used for the direct sulphate

forcing shown in Fig. 1d. The indirect forcing pattern is quite different to the direct pattern, with a distinct land–sea contrast. Since additional aerosols have the greatest impact in the cleaner southern hemisphere, the inter-hemispheric difference is significantly smaller than for the direct forcing (Chuang et al., 1997; Feichter et al., 1997). For sulphate aerosol, the direct forcing is around four times higher in the northern hemisphere than in the southern, while the value for the indirect forcing may be as low as two times higher. For the same reason, the forcing may be significantly greater over the North Atlantic than the North American continent (Chuang et al., 1997; Feichter et al., 1997; Jones and Slingo, 1997).

2.6. Other forcing mechanisms

Other radiative forcing mechanisms have been proposed; many were discussed, although not quantified, in early climate change reviews (e.g., SMIC, 1971; Matthews et al., 1971).

Changes in surface reflectance can result from changes in land use (Henderson-Sellers, 1995). It has been estimated (Hansen et al., 1997b) that the global-mean forcing due to this effect is -0.4 W m^{-2}, about half of which is estimated to have occurred since the industrial revolution. The biggest effect was found at high latitudes as snow-covered forest has a lower reflectance than snow-covered areas that have been deforested. Given the lack of detailed recent studies, we take -0.2 W m^{-2} as a central estimate, with at least a factor of two uncertainty and a 'very low' confidence.

Stratospheric water vapour concentrations are believed to increase in response to increased methane concentrations, as water vapour is a product of methane oxidation. IPCC (1990) originally proposed that this water vapour forcing was one third of the size of the methane forcing itself but this figure has been contested (IPCC, 1994). Mid-latitude balloonsonde ascents since 1981 have indicated an upward trend of 0.5 to 1% per year in the lower stratosphere (Oltmans and Hofmann, 1995); this has been calculated (Forster and Shine, 1997) to cause a local forcing of $+0.06$ W m^{-2} $decade^{-1}$. Evans et al. (1998) have also found evidence of an increase in stratospheric water vapour from satellite observations, albeit for a shorter period (1992–1996) and at higher levels (30–65 km). Since the global extent and the duration of this change are unknown, the forcing is very uncertain. In addition, part of this water vapour change may be natural or a response to a climate change (for example, by changing the amount of water vapour transported through the tropical tropopause) and so would constitute a climate feedback rather than a forcing (see Appendix A). We thus refrain from making a global estimate.

The concentrations of other gases in the atmosphere, such as non-methane hydrocarbons, have also been perturbed by human activity and can reach ppbv concentrations in industrial regions; while their effect on tropospheric ozone has been considered in many of the calculations of that forcing, they also can be, on a per molecule basis, strong greenhouse gases. Highwood et al. (1999) attempted a preliminary estimate of the direct forcing due to increased concentrations of 15 such hydrocarbons, and tentatively concluded that the global-mean forcing was small, of order 0.01 W m^{-2}.

The potential climatic effect of aircraft emissions has long been of concern (e.g., SMIC, 1971; Matthews et al., 1971) and is the subject of renewed interest (e.g., Brasseur et al., 1998a). Aircraft emissions of carbon dioxide and oxides of nitrogen (which impact on ozone) are generally implicitly considered in calculations of the well-mixed greenhouse gases and tropospheric ozone change. Water vapour emissions appear to be of negligible climatic importance (Ponater et al., 1996; Rind et al., 1996). Contrail formation may have changed the nature and occurrence of upper tropospheric clouds. One estimate is that within the busiest flight routes, contrails have added around 0.005 (with an upper limit of 0.02) to the fractional cloudiness (Ponater et al., 1996); this has been estimated to lead to local perturbations of between 0.1 and 1 W m^{-2} and around an order of magnitude less for the global change. Such estimates are made difficult due to the fact that changes in upper tropospheric clouds cause radiative forcings of opposite sign in the solar and thermal infrared leading to uncertainty in the sign of the net impact. Wyser and Ström (1998) have investigated the possible impact of aircraft emissions on the crystal sizes of natural cirrus; they found that the sign of the

impact depended on as yet poorly characterised crystal sizes in the natural cirrus.

3. Relationship between radiative forcing and climate response

If Eq. (1) in Appendix A is to approximately model the global-mean surface temperature change, not only must it perform for the diverse set of spatial distributions of forcing shown in Fig. 1, it must also cope with the diverse ways in which the mechanisms impact on the radiation budget. For example, when carbon dioxide is changed, the forcing is initially felt mostly in the upper troposphere rather than at the surface. For stratospheric ozone depletion, the surface is heated by the increased solar radiation, while the changes in the thermal infrared cool the upper troposphere. For increases in both soot and sulphate aerosols, the surface is deprived of solar radiation; in the case of soot, this is more than compensated for by the increased absorption within the troposphere, while for sulphate the whole surface-troposphere system loses energy.

Normally, a direct demonstration of the link between forcing and response is difficult due to the gradual decadal scale changes in forcing and a similarly slow timescale in the response. However, the eruption of Mount Pinatubo in 1991 caused such a sudden increase in stratospheric aerosol loading that the forcing (which reached -4 W m^{-2}) was clearly detectable from satellites (Minnis et al., 1993; IPCC, 1994). The resulting tropospheric cooling was predicted well by a GCM (Hansen et al., 1992, IPCC, 1994) and this has helped increase confidence in the ability of models to respond correctly to radiative forcings. This case also demonstrated the complexity of climate response; superimposed on the global-mean cooling was a dynamically induced warming of the continents during winter, which has been attributed to the effects of the volcanic aerosol (Graf et al., 1993; Robock and Mao, 1995). GCM studies (e.g., Cox et al., 1995; Erickson et al., 1995; Mitchell et al., 1995) emphasise that the geographical patterns of forcing and response can be distinctly different; this warns against simple correlation of patterns of radiative forcing (such as those in Fig. 1) with patterns of observed climate change. On the other hand, several studies (e.g., Engardt and Rodhe, 1993; Hunter et al., 1993; Karl et al. 1995b) have noted the existence of a correlation between aerosol forcing patterns and surface temperature response; it is not clear whether this correlation is coincidental, or an indication of a stronger connection between local forcing and local response than is implied by the GCMs.

The relationship between radiative forcing and climate response has been explored using systematic experiments with GCMs; these models allow a consistent interaction between radiative forcing and atmospheric circulation. Cox et al. (1995) compared their GCM's response to an increase in carbon dioxide with the response to an increase in sulphate aerosol which was tuned to give a global-mean forcing of the same size but opposite sign to the carbon dioxide forcing. When both forcings were included, the global-mean response was only 0.01 K, showing a remarkable degree of compensation; for each forcing individually, the global-mean surface response would have been about 0.6 K. However there were more significant hemispheric-scale temperature changes (of order 0.1 K) and regional-scale responses (exceeding 1 K in some regions). Ramaswamy and Chen (1997) compared the response for an increase in carbon dioxide with a set of radiative forcings obtained by perturbing low cloud properties, on both global and continental scale. To within $\pm 15\%$, the global-mean climate sensitivity was independent of the nature of the perturbation, although, again, the local response differed among the experiments. Haywood et al. (1997c) have shown a remarkable degree of additivity in the climate response when different forcings are applied; the sum of the climate response (e.g., surface temperature changes and precipitation changes) for greenhouse gases and sulphate aerosol changes applied separately, was found to be similar to the response when both forcings were added simultaneously.

The most comprehensive study of the forcing–response relationship is that of Hansen et al. (1997a). Their simplified GCM (it represented 120° of longitude rather than 360°, with nine levels in the vertical) had a relatively low computational demand; so compared with any other study, they explored a very large range of perturbations.

In common with other GCM studies (Cox et al., 1995; Ramaswamy and Chen, 1997), Hansen et al. (1997a) comment that their results 'reaffirm the value of radiative forcing' but they also indicate the care with which the concept must be used. For the more realistic forcings used, and when clouds were not allowed to respond to climate change, the global-mean climate sensitivity varies by about $\pm 30\%$; when cloud feedbacks were included, the sensitivity varies by about $\pm 50\%$. Mechanisms that caused the departure of the climate sensitivity from a constant value were identified. These included processes that altered the lapse rate of temperature in the upper troposphere, or led to changes in the local heating rate that altered cloudiness. Given the notorious difficulties in modelling clouds in climate models, results from a single model need to be treated with some caution. Nevertheless, if future work with a range of GCMs confirm these results, it may be possible to label forcings by the way they impact on the climate sensitivity and to construct 'effective' radiative forcings which would allow forcings due to different mechanisms to be compared more meaningfully.

One result from the Hansen et al. (1997a) study is of particular note. It was found that absorbing aerosols, because they warm cloud layers and decrease cloudiness, were more effective than indicated by their radiative forcing. An important practical consequence is that the amount of aerosol absorption required to change an aerosol's net climatic effect from cooling to warming is less than that indicated by radiative forcing, and closer to the amount of absorption that is typically observed. This result led Hansen et al. (1997b) to conclude that the direct forcing due to increased amounts of aerosols may actually be close to zero.

Simpler climate models have also been used. A two-dimensional (latitude–height) study (Mackay et al., 1997) with a simple representation of atmospheric and oceanic heat transport found the climate sensitivity for stratospheric ozone depletion to be around three times higher than that for well-mixed greenhouse gases, apparently because of more efficient upper tropospheric heat transport in the ozone case. This difference in sensitivity is much higher than the 30% difference found in similar experiments using a GCM (Hansen et al., 1997a); studies with more sophisticated models would be required to check the robustness of this result.

One-dimensional (vertical) radiative–convective models formed the basis of the original development of the concept of radiative forcing. One such model (Forster et al., 1997) has been used to investigate the dependence of the forcing–response relationship on the definition of tropopause position (see Appendix A). Tropopause definitions are somewhat arbitrary, but in radiative–convective models a natural definition is the altitude at which the atmosphere changes from being in radiative–convective equilibrium (and hence, most tightly coupled to the surface) to being in pure radiative equilibrium. For forcings with a complex vertical structure, such as stratospheric ozone depletion, the appropriate choice of tropopause definition was shown to be important if the climate sensitivity was to remain quasi-constant. Although the definition of the appropriate tropopause level is much more difficult in a GCM (and in the real atmosphere), this study indicated that care needs to be taken in GCM studies; departures from the expected forcing–response relationship may arise from an inappropriate choice of the altitude at which the radiative forcing is calculated.

4. Conclusions

Fig. 2 is our update of a figure in IPCC (1995) showing global-mean radiative forcings, together with an indication of the size of natural forcing mechanisms; for comparison, we also include estimates of the forcing due to changes in solar output, and the peak inter-decadal impact of volcanic eruptions—details are given in the caption. On a global-mean basis, and on current understanding, the radiative forcing due to the well-mixed greenhouse gases seems to dominate over the other forcings, but until the size of the indirect forcing is more firmly established, the extent of this dominance remains uncertain. The geographical patterns in Fig. 1 emphasise that the global mean is a drastic simplification of the entire radiative forcing picture. Fig. 1 also shows some intriguing juxtapositions of patterns. For example, over North Africa, the high surface reflectance, high surface temperature and small amount of cloud combine in such a way that causes a distinct regional

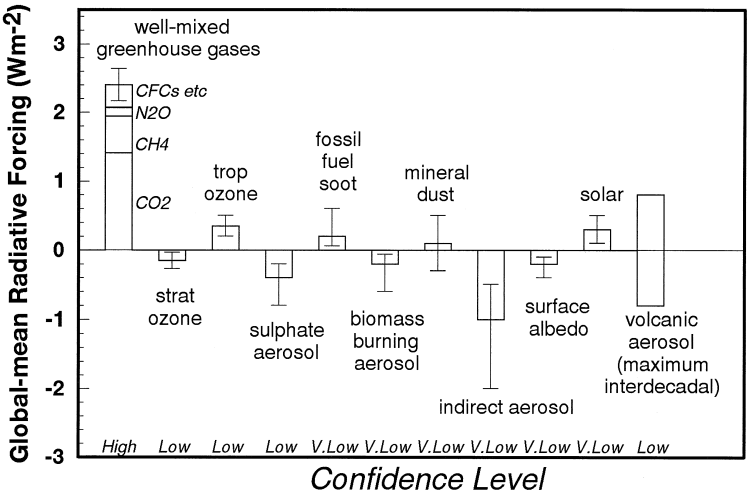


Fig. 2. The global-mean radiative forcing due to individual mechanisms for the period from pre-industrial times to the present. The 'error bars' indicate the range of forcings inferred, where possible, from the recent literature. The confidence level indicates our confidence that the forcings actually lie within these error bars; it is rather subjectively based on the degree to which the constituent distributions used in the forcing calculations can be verified, the difficulties in modelling the radiative effect of that constituent and the sophistication of the methods used to derive the forcing. The solar forcing is taken from IPCC (1995) and the volcanic forcing (Sato et al., 1993) is shown to illustrate the difference in decadal-mean forcings between a decade which has had several climatically significant eruptions (such as 1983–1992) and a decade with little such activity (such as 1950–1959). Any cancellation between mechanisms in terms of global-mean forcing must not be taken to indicate a reduction in climate impact, as regional scale changes can still be significant.

peak in positive radiative forcing due to tropospheric ozone, fossil-fuel soot, soil dust and, to a lesser extent, the well-mixed greenhouse gases. Over oceans, the negative radiative forcing due to the direct and indirect effect of sulphate aerosols and the soil dust superimpose. A GCM study would be required to examine the consequences of these juxtapositions on climate response.

Although many mechanisms by which human activity can impact on climate have long been identified, it is only in the past few years that quantification of many of these has been attempted. Many uncertainties remain, but the progress has been encouraging. Key variables which affect the size and sign of the forcing have been identified and there has been a rapid development of CTMs to derive pre-industrial and present day fields of poorly-observed constituents. These models will become more sophisticated and include more components; this, together with the availability of results from different modelling groups which allows critical comparison, should lead to significant further progress. Such studies will need to become more integrated; to date, they have generally considered each aerosol component, or stratospheric and tropospheric ozone, independently. The changes in these constituents are sufficiently interlinked that interactions between them may cause significant revisions to present results. In addition, most studies have concentrated on the total change since pre-industrial times, rather than the way that the forcing has evolved with time; different forcings will probably have evolved in different ways, with consequences for the evolution of the climate response. It is essential that the observational data for the critical evaluation of the CTMs is expanded, in order to strengthen confidence in the ability of CTMs to model constituent fields.

The evaluation of these models is often severely limited by the lack of good observational data but the local size of, for example, aerosol effects means that they are measurable in field experiments (Russell et al., 1997). A number of major field campaigns have recently occurred, or are being planned, including the Aerosol Characterization Experiments (e.g., Bates et al., 1998) and the Tropospheric Aerosol Radiative Forcing Observational Experiment (e.g.,

Hegg et al., 1997); these are measuring the aerosol characteristics and radiative effects in a variety of clean and polluted environments and promise to provide major advances in our understanding. A further step is required for radiative forcing calculations, as long-term monitoring is crucial (Hansen et al., 1995; Karl et al., 1995a); it needs to be ensured that adequate surface and space-based monitoring systems, which have not been available in the past, are put in place.

This review has shown that, in addition to the well-mixed greenhouse gases, there are several other forcing mechanisms (both natural and related to human activity) which each, given current uncertainties, may have roughly similar sizes to each other, but vary in sign and spatial pattern. Several recent studies (Mitchell et al., 1995; Santer et al., 1996; Tett et al., 1996; Hasselmann, 1997) have used model-derived patterns of climate response in attempts to attribute observed climate change to human activity. They have, however, at most, included the forcings due to the well-mixed greenhouse gases, the direct effects of sulphate aerosols and stratospheric ozone. Given the uncertainties in the latter two forcings, and the likely presence of several other forcings of similar size but differing sign and geographical pattern, their conclusions may not be robust when more mechanisms are added. One obvious and simple illustration is that the good agreement between observed and modelled global-mean surface temperature changes (Mitchell et al., 1995, Haywood et al., 1997c) would be significantly modified if, for example, the additional negative forcing due to the indirect effect of sulphate aerosols had been included; these studies may have, fortuitously, obtained an approximately correct net forcing when including only the well-mixed greenhouse gases and direct sulphate forcings.

Despite these caveats about the results of these attribution studies, we must conclude that human activity is having a marked and complex impact on the Earth's radiation budget. Even if the details of the resulting climatic effect remain uncertain, we would expect a significant response to have already occurred. Given the rapid development of our understanding of radiative forcing mechanisms in the past decade or so, we cannot be confident that all the major mechanisms have yet been identified.

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Appendix A. Radiative forcing

Radiative forcing provides a convenient first-order measure of the climatic importance of perturbations to the planetary radiation balance. For small perturbations, climate model experiments (IPCC, 1994; Hansen et al., 1997a) indicate that the equilibrium global-mean surface temperature change (ΔT_s) is approximately linearly related to the radiative forcing (ΔF) (if imposed for a time period long compared to the multi-decadal response time of the climate system) so that

$$\Delta T_s \approx \lambda \Delta F \quad (1)$$

where λ is the climate sensitivity (in K (W m^{-2})$^{-1}$). The value of λ is determined by the direct response of the surface temperature to the change in energy balance and by a number of feedbacks. The feedbacks, which arise as part of the response to climate change, include changes in water vapour, sea-ice and snow extent, and cloud properties. In this conceptual model, the radiative forcing is explicitly separated from the climate response; it is important to note that while water vapour is the dominant greenhouse gas, its role in climate change is via a feedback, as its tropospheric concentrations are largely controlled by moist physical processes and atmospheric circulation.

Results from early climate model experiments indicated that, if appropriately defined (see below), an important feature modelled by Eq. (1) is that the climate response depends on the size, but not the nature, of the global-mean radiative forcing; thus a change in solar output, which alters only the absorbed solar radiation, would have the same impact as a change in carbon dioxide concentration (where the effect is mainly in the thermal infrared) if they cause the same radiative forcing.

The conceptual model of radiative forcing views the surface–troposphere system as being a single thermodynamic system, coupled by radiative and

convective energy exchanges. Hence, radiative forcing is more strictly defined as the change at the tropopause. One significant elaboration of this definition concerns the role that stratospheric temperature changes have on the surface–troposphere energy balance. If the atmosphere is perturbed by, for example, altering the carbon dioxide concentration, the stratosphere responds within a few weeks, much faster than the decadal timescale for the surface–troposphere system response, which is controlled by the thermal inertia of the oceans (e.g., IPCC, 1994; Hansen et al., 1997a). Hence, changes in the infrared emission from the stratosphere to the troposphere, as a result of a stratospheric temperature change, are taken to be part of the forcing. Without inclusion of this 'stratospheric adjustment' in the calculation of ΔF, the value of λ in Eq. (1) is much more variable among different climate change mechanisms (Ramanathan et al., 1987; IPCC, 1994; Forster et al., 1997; Hansen et al., 1997a). For ozone change in the lower stratosphere, this adjustment is crucial for determining the sign of the climate response. It is important at the 10–20% level for changes in carbon dioxide and tropospheric ozone changes, but of little importance for tropospheric aerosol changes.

One important reason for using radiative forcing rather than the resultant change in climate is that there is currently a large uncertainty (e.g., IPCC, 1995) in the value of λ, with different models giving values in the range from about 0.3 to 1.1 K (W $m^{-2})^{-1}$; most of this uncertainty is due to model differences in cloud feedbacks. Hence, for a particular climate change mechanism, it is more straightforward to compare different estimates of ΔF than ΔT_s. There are other reasons for calculating radiative forcing. It is computationally much less demanding than using a sophisticated climate model; the sensitivity to poorly known parameters can be more easily explored, and more sophisticated radiative transfer models can be used. Radiative forcing is also an important tool for climate modellers to assess whether any proposed mechanism justifies a computationally expensive climate model calculation; indeed, it is an important tool for monitoring model performance, as dramatic departures from the response anticipated from Eq. (1) can indicate faults in models.

A number of caveats should be noted. First, the concept was developed before the detailed consideration of the current wider range of climate change mechanisms. Its applicability to these mechanisms is under active investigation, as discussed in Section 3; indications so far are that it remains a useful, if imperfect, measure. Secondly, it must be emphasized that Eq. (1) refers to global means. If different climate change mechanisms, with radiative forcings of opposing signs, were to fortuitously combine to give a small global-mean forcing, significant regional-scale climate changes are still likely, even if the global-mean surface temperature change obeyed Eq. (1).

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