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CHAPTER 7

Radiative Forcing of Climate

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Chapter 7

Radiative Forcing of Climate

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SCIENTIFIC SUMMARY

Global Warming Potentials

Direct Effect. Direct global warming potentials (GWP) have been recalculated for tropospheric, well-mixed, radiatively active species (CH_4 , N_2O and the halocarbons). We use updated lifetimes of these species and follow the same methodology as in the Intergovernmental Panel on Climate Change report (IPCC, 1990), for time horizons corresponding to 20, 50, 100, 200, and 500 years. The new GWP results include a correction in the values for methane arising due to a typographical error in the IPCC (1990) report, as well as changes in values for other species that are a manifestation of the updated lifetimes. However, there still exist uncertainties in these calculations due to uncertainties in the carbon cycle.

Indirect Effect. While chemical reactions involving the radiatively active atmospheric species can contribute to the GWPs, our ability to estimate them is restricted at present owing to complexities in the chemical processes and uncertainties in the temporal and spatial variations of various species. While the sign of the radiative forcing due to some of the indirect effects can be evaluated with a fair measure of confidence (Chapter 5), the quantitative aspects of the indirect effects are more difficult to ascertain than was anticipated earlier in IPCC (1990) and we do not recommend the use of those values.

Radiative forcing due to trace gases, including ozone (1979–1990)

Stratospheric Ozone. The observed global ozone losses in the lower stratosphere cause a significant forcing of the climate system. The physical effects due to this loss consist of an increase in the solar and a decrease in the longwave forcing of the surface-troposphere system, together with a tendency to cool the lower stratosphere. The latter effect amplifies the longwave influences and can thereby give rise to a significant net negative radiative forcing of the surface-troposphere system, but the magnitude of the induced forcing is very sensitive to the change in the lower stratospheric temperatures. This tendency is opposite to the direct positive (greenhouse) forcing exerted by the well-mixed gases and is pronounced in the mid-to-high latitudes of both hemispheres, and during all seasons. In fact, if the tendency to cool the lower stratosphere is fully realized as equilibrium temperature change, the magnitude of the negative ozone radiative forcing in the mid-to-high latitudes can be larger than the decadal greenhouse forcing due to the CFCs over this period, and could also be a significant fraction of the total decadal greenhouse forcing due to the non-ozone gases.

Tropospheric Ozone. Ozone in the troposphere, although present in smaller amounts than that in the stratosphere, has a significant opacity and exerts a greenhouse effect (positive forcing), with the sensitivity being greatest for ozone changes in the upper troposphere. However, the databases for tropospheric ozone trends are sparse and therefore inadequate for quantifying the global radiative influences due to changes in tropospheric ozone.

Radiative forcing due to sulfate aerosols

Tropospheric Aerosols. Fossil fuel emissions over the past century have increased significantly the tropospheric sulfate aerosol concentrations. The contribution of this species to the direct clear-sky radiative forcing of the Northern Hemisphere is opposite to that due to the greenhouse gases and is estimated to be a significant fraction of the trace gas forcing. Although we are confident of the sign of the forcing, uncertainties exist owing to the spatial inhomogeneity of the aerosol distributions and a lack of understanding of the aerosol effects in cloudy regions.

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Stratospheric Aerosols. With the eruption of Mt. Pinatubo in mid-1991, there is again a radiative forcing of the climate system due to increases in the concentrations of the stratospheric sulfate aerosols. These aerosols produce a net negative radiative forcing of the climate system which is comparable to the positive ones due to the greenhouse gases, but this forcing is short-lived. The presence of the aerosols in the stratosphere also contributes to a warming tendency in the lower stratosphere. Preliminary observations indicate about a 4 K warming of the tropical lower stratosphere two months following the Mt. Pinatubo eruption.

7.1 INTRODUCTION

This chapter is an update of the scientific discussions presented in Chapter 2 of the Intergovernmental Panel on Climate Change report (IPCC, 1990; hereafter referred to as IPCC) concerning the atmospheric radiative and chemical species of significance for climate change. There are two major objectives of the present update. The first is an extension of the discussion on the Global Warming Potentials (GWPs), including a reevaluation in view of the updates in the lifetimes of the radiatively active species. The second important objective is to underscore major developments in the radiative forcing of climate due to the observed stratospheric ozone losses occurring between 1979 and 1990. The contents of this chapter are in the following sections:

- 7.2 Definitions of radiative forcing,
- 7.3 GWPs of the well-mixed trace gases using new lifetimes; also included is the sensitivity of the radiative forcing to the absolute concentrations of greenhouse gases, and the radiative forcing due to increase of water vapor in the stratosphere resulting from the oxidation of methane,
- 7.4 Radiative forcing due to the increases in non-ozone trace gases (1979–1990),
- 7.5 Characterization of the radiative forcing due to changes in atmospheric ozone that have occurred over the past decade (1979–1990),
- 7.6 Radiative forcing induced by increases in tropospheric sulfate aerosol concentrations, and
- 7.7 Stratospheric aerosol effects following the eruption of the Mt. Pinatubo volcano.

Several radiative transfer models have been employed to obtain the results described here. Those that performed calculations exclusively for this study are listed in Table 7-1.

7.2 RADIATIVE FORCING

The radiative forcing due to a perturbation in the concentration of a gas is defined by the net radiative flux change induced at the tropopause. The forcing is usually interpreted as a gain (positive) or a loss (negative) for the surface-troposphere system as a whole. The rationale for this concept arises from exercises with one-dimensional radiative-convective models where the change in the surface temperature can be

Table 7-1 Radiative transfer models employed in this study. Models I and III have a spectral resolution of 10 cm^{-1} . Model V includes a one-dimensional chemistry model as well.

Model	Institution	References
I	Geophysical Fluid Dynamics Laboratory	Fels <i>et al.</i> (1980); Mahlman and Umscheid (1984); Ramaswamy <i>et al.</i> (1990)
II	Geophysical Fluid Dynamics Laboratory	Ramaswamy and Ramanathan (1989)
III	University of Reading	Shine (1991)
IV	State University of New York at Albany	Wang and Molnar (1985); Fisher <i>et al.</i> (1990)
V	Lawrence Livermore National Laboratory	Miller <i>et al.</i> (1991)

related simply to the net radiative flux change at the tropopause (WMO, 1986). This has led to the adoption of the surface-troposphere system as a convenient means to obtain a perspective of the trace gas radiative effects. The instantaneous radiative forcing is given by the change obtained keeping all parameters (including temperature and water vapor) fixed in both the troposphere and the stratosphere, with the meteorological conditions conforming to an assumed atmospheric state.

A second definition of radiative forcing, emerging from the discussions in WMO (1986) and IPCC relates to the case when the stratospheric temperatures are allowed to relax to a new equilibrium under the assumption of a constant dynamical heating—the so-called Fixed Dynamical Heating (FDH) concept (Fels and Kaplan, 1975; Ramanathan and Dickinson, 1979; Fels *et al.*, 1980). This also constitutes a generalization of the global mean concept of relaxing the stratosphere back to a radiative equilibrium (Hansen *et al.*, 1981), with the proviso that the tropospheric temperatures as well as the water vapor everywhere remain unchanged, being the same as in the unperturbed state. This definition makes use of the fact that the model stratosphere adjusts or comes to equilibrium more rapidly (usually within 60 to 90 days) to the perturbation induced by a change in a specific trace

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gas concentration than do the troposphere and the surface (several years to possibly several decades).

For the purposes of this assessment, we will term the *instantaneous* forcing resulting from the first definition as *Mode A*, and that from the second definition involving the *stratospheric temperature adjustment* as *Mode B*. It is the Mode B result that describes more appropriately the long-term (seasonal and longer time scales) forcing of the surface-troposphere system (WMO, 1986; IPCC, 1990). Further, Mode A can be interpreted as the case without temperature feedback in the stratosphere while, in Mode B, there is a temperature feedback in the stratosphere. In both modes, tropospheric feedback mechanisms are excluded.

7.3 GLOBAL WARMING POTENTIAL

7.3.1 Definition

As a relative measure of the possible warming effect of the surface and the troposphere arising due to the emissions of each greenhouse gas, IPCC presented the concept of the Global Warming Potential. The GWP of a greenhouse gas takes into account the radiative forcing due to a single pulsed emission of that gas, as well as its lifetime within the atmosphere. It is thus related to the emissions of greenhouse gases and their lifetimes. Since IPCC discusses the GWP concept in considerable detail, only the salient features of the GWP are emphasized here.

The GWP of a well-mixed gas is defined (IPCC, page 58) formally as the time-integrated commitment to radiative forcing from the instantaneous release of 1 kg of a trace gas expressed relative to that from the release of 1 kg of CO₂. This definition requires a consideration of the following factors:

- choice of a reference molecule and its lifetime, as well as the lifetimes of other gases,
- the radiative forcing due to a change in the concentration of a gas,
- the time horizon over which the radiative forcings have to be integrated,
- the background concentrations of various species, and the atmospheric temperature and moisture profiles, and
- the indirect effects due to chemical reactions that are accompanied by changes in the concentrations of the radiatively active species.

The GWP concept is derived from the globally- and annually-averaged net radiative flux change at the tropopause. Accordingly, the surface-troposphere radiative forcing employed in GWP determinations is obtained from a radiative transfer model for global mean atmospheric conditions. The principal significance of the GWP concept lies in the fact that it offers a simple yet reasonable characterization of the relative global mean radiative effects due to changes in the concentrations of the well-mixed species. This renders it particularly useful in the context of policy-making decisions concerning the emissions of the well-mixed greenhouse gases. However, as noted subsequently (section 7.3.8), this concept does have serious limitations.

7.3.2 Reference Molecule and its Lifetime

Given the conceptual framework of the GWP and its implications for policy-making, the choice of a reference molecule is dictated by the need for simplicity and the need to evaluate the results in terms of the dominant contributor in the greenhouse gas problem. IPCC chose CO₂ as the reference molecule for the GWP determinations. Although another gas or surrogate would have a simpler atmospheric decay rate compared to CO₂ (e.g., CFCs; see Fisher *et al.*, 1990) and can equally well be chosen as the reference, the evaluation of GWPs presented here continues to use CO₂ as the reference gas. The chief reason for this is the importance of CO₂ as the gas of primary concern to future climate change. Further, again following IPCC and notwithstanding the uncertainties in the carbon cycle, the results from the Siegenthaler (1983) model are employed to estimate the lifetimes of CO₂.

7.3.3 Time Horizons for GWPs

Since greenhouse gases differ in their relative reactivity and their sink mechanisms, they remain in the atmosphere for different time periods, *i.e.*, they have different lifetimes. The calculation of GWPs, thus, depends on the time-period of integration chosen for the analysis. There is, however, no given value of integration time for determining GWPs that is ideal over the range of uses of this concept.

The choice of a time-scale for integration of the GWP calculation, however, need not be totally arbi-

trary. The risks of climate change are associated with different kinds of undesirable changes. For example, rapid climate changes over successive decades may have adverse impacts to biological systems as climate zones shift more quickly than natural systems can migrate. Likewise, a several degree increase in surface temperature over the next century or two, a somewhat longer time horizon, could have adverse effects on both human communities and natural ecosystems. Each type of system has its own characteristic thresholds of sensitivity to different types of damage. The choice of the yardstick for measuring the risks of climate change depends on the type of undesirable changes that are of greatest interest to the analyst or policymaker.

Although most of the public attention to the climate problem has been concentrated on one indicator of risk (*i.e.*, the global average surface temperature change), several other indicators of unwanted change, too, could have important climate change ramifications. Also, they could be of importance from the policymaker's perspective. The choice of which indicator to use (and which time horizon to employ) for the GWP analyses is determined by the type of undesirable change that is under consideration.

The effect of the surface inertia of the climate system associated with the world's oceans influences the time-scales associated with a given climatic response and the resulting impacts. Within the climate system, the exchange of energy between the oceans and the atmosphere results in a lag in the temperature response to a given forcing. The lag in the climate response is estimated to be between 10 and 100 years (IPCC).

Several possible indicators of change (and the associated time horizons for analyses of GWPs) are shown below. The column on the right illustrates the characteristic integration periods that would be appropriate to capture the important aspects of these different indicators:

Climate Change Indicator	Integration Time Period
Maximum change in temperature	~100 years
Rate of change in temperature	~ 20-50 years
Maximum change in sea level	> 100 years
Rate of change in sea level	> 50 years

GWPs in this report are calculated over time horizons of 20, 50, 100, 200, and 500 years (as compared to 20, 100, and 500 years only in IPCC). These

five different time horizons ought to provide a useful set of reference values for policy decisions. As suggested above, the GWPs estimated for long time periods provide a measure of the cumulative chronic effects on climate. Integrations to extremely long times (*i.e.*, 500 years) are subject to significant uncertainties in the decay rate of atmospheric CO₂. On the other hand, the 20 and 50-year integrations are representative of the time scales for the maximum rate of response of temperature.

The GWPs evaluated over the 100-year period appear generally to provide a balanced representation of the various time horizons for climate response. This is a time scale that includes due consideration of the ocean thermal inertia and its impacts on the global mean surface temperature. In addition, carbon cycle models also indicate that this time period broadly represents the time scale over which a significant fraction of CO₂ is removed from the atmosphere. However, policy analyses emphasizing the rate of temperature change or the rate of sea level rise may find the 50-year integration period to provide a better representation of the climate responses. Considerations of the shortest time scale (*i.e.*, 20 years) suggest one way to reduce the rate of increase of the radiative forcing, particularly for short-lived species.

7.3.4 Direct GWPs of Well-Mixed Trace Gases

Based on the above considerations, and employing the updated lifetimes of the various species (Chapter 8), and the radiative forcings given in Table 2-3 of IPCC, new direct (*i.e.*, in the absence of atmospheric chemistry considerations) GWPs of several well-mixed species are determined. These are listed in Table 7-2 for the five time horizons and may be compared with Table 2-8 in IPCC. The forcings employed here conform to the Mode B definition discussed in Section 7-2, thus implicitly allowing for the temperature feedback in the stratosphere due to the particular greenhouse gas. Changes in the lifetime and variations of radiative forcing with changes in the background concentrations of species are neglected.

The direct GWPs for methane in Table 7-2 are substantially higher than those inferred from IPCC owing to a typographical error in that report. The

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Table 7-2 Direct global warming potentials of several well-mixed trace gases relative to CO₂. The GWPs of the various non-CO₂ species are calculated for each of five time horizons (20, 50, 100, 200 and 500 years) using, as in IPCC, the carbon cycle model of Siegenthaler (1983). (Note that IPCC contained a typographical error which led to incorrect values for the direct GWP of methane.)

Gas	Lifetime (years)	Time Horizons				
		20 years	50 years	100 years	200 years	500 years
CO ₂	#	1	1	1	1	1
CH ₄	10.5	35	19	11	7	4
N ₂ O	132	260	270	270	240	170
CFC-11	55	4500	4100	3400	2400	1400
CFC-12	116	7100	7400	7100	6200	4100
HCFC-22	15.8	4200	2600	1600	970	540
CFC-113	110	4600	4700	4500	3900	2500
CFC-114	220	6100	6700	7000	7000	5800
CFC-115	550	5500	6200	7000	7800	8500
HCFC-123	1.71	330	150	90	55	30
HCFC-124	6.9	1500	760	440	270	150
HFC-125	40.5	5200	4500	3400	2200	1200
HFC-134a	15.6	3100	1900	1200	730	400
HCFC-141b	10.8	1800	980	580	350	200
HCFC-142b	22.4	4000	2800	1800	1100	620
HFC-143a	64.2	4700	4500	3800	2800	1600
HFC-152a	1.8	530	250	150	89	49
CCl ₄	47	1800	1600	1300	860	480
CH ₃ CCl ₃	6.1	360	170	100	62	34
CF ₃ Br	77	5600	5500	4900	3800	2300

#(see Table 2.8, IPCC, 1990)

other differences between Table 7-2 and Table 2-8 of IPCC are a manifestation of the changes in the lifetimes. Species whose lifetimes have changed substantially include F-125 (45 percent higher) and F-141b (35 percent higher), F-143a (57 percent higher) and CF₃Br (30 percent less). The new direct GWPs in Table 7-2 are generally within 20 percent of the IPCC values. The exceptions are the above-mentioned species viz., F-125, F-141b and F-143a, all of which have an increase in GWP exceeding 20 percent for the 100- and the 500-year time horizons, while CF₃Br has a decrease of more than 20 percent for the 500-year horizon.

7.3.5 Sensitivity of Radiative Forcing to Background Trace Gas Concentrations

One important parameter determining the radiative forcing due to changes in the concentrations of trace

gases is the background concentrations of the trace species themselves. The effect of this parameter is investigated here using one-dimensional radiative-convective models. Specifically, the perturbation in the radiative fluxes is examined for two different background concentrations, one corresponding to present (1990) atmosphere and the other to a postulated 2020 atmospheric profile (Chapter 8). Each atmosphere is brought to a model-determined radiative-convective equilibrium. To each of these atmospheres in equilibrium, an identical perturbation is applied corresponding to an increase in the concentration of a particular trace gas, keeping the concentrations of the other species fixed. The perturbations are evaluated for the Mode A condition, *i.e.*, without the stratospheric temperature feedback.

Models I, II, and IV performed computations for specified perturbations (see Table 7-3) in the concentrations of CO₂, CH₄, N₂O, F-11 and F-12. The results,

listed in Table 7-3, are quoted as the ratio by which the radiative forcing in 2020 changes with respect to that in 1990. The perturbations applied are arbitrary and correspond to changes in the concentrations of each trace gas occurring between 1980 and 1990 (Chapter 8).

The results from the various models agree well with each other and are completely understandable in terms of the nonlinearities in the radiative properties of the individual species (IPCC). The decrease in the forcing due to CH₄ and N₂O is less than that for CO₂, so that the GWPs of CH₄ and N₂O would be greater in 2020 than in 1990. The changes for F-11 and F-12 are negligible. These examples confirm the dependence of the GWPs on the assumed background atmospheric concentrations of the trace species.

Table 7-3 Ratio of the Mode A radiative forcings due to an increase in the concentration of a trace gas, as computed for the atmospheric profiles in the years 2020 and 1990. The atmospheres correspond to those obtained from a one-dimensional radiative-convective model, with the background concentrations of the trace gases in 1990 and 2020 following Chapter 8. The perturbation applied to the 1990 and 2020 atmospheres in radiative-convective equilibrium are listed below. The ratios are derived from Models I, II and IV, respectively.

Gas	Perturbation applied	Model I	Model II	Model IV
CO ₂	17 ppmv	0.85	0.86	0.87
CH ₄	15 ppmv	0.89	0.87	0.95
N ₂ O	8 ppbv	0.95	0.92	0.94
CFC-11	111 pptv	1.0	1.0	1.0
CFC-12	170 pptv	1.0	1.0	1.0

7.3.6 Indirect GWP Effects

While a sizable body of knowledge is available to indicate the sign of the indirect effects as induced by chemical processes, the magnitudes of these effects are still not easily quantified without considerable uncertainties. Chapter 5 presents estimates of the changes in species concentrations induced by chemical interactions. Some of these could contribute to the GWP of certain species listed in Table 7-2. We recognize that these indirect effects are more difficult to estimate than the direct effects owing to the complexity of the chemi-

cal reactions and the temporal and spatial dependence of the involved species' concentrations. While the sign of the indirect effects can be estimated with a fair degree of confidence (Chapter 5), the quantitative aspects merit further detailed investigations. In fact, the values for the indirect effects in IPCC could be very uncertain and their use is not recommended. It is now known that the value for NO_x, in particular, may have been overestimated substantially in the IPCC report (Johnson *et al.*, 1992). It is possible that, for methane, some progress in quantifying the uncertainties in the indirect effects may be possible in the near future. As an example, the indirect enhancement in the methane radiative forcing due to a specific chemical transformation is discussed in the next section.

7.3.7 Indirect GWP of Methane Due to Oxidation to Water Vapor

As noted in IPCC, the oxidation of methane in the stratosphere results in an increase in the water vapor content at those altitudes, which, in turn, contributes to the greenhouse effect. In this study, a one-dimensional chemistry algorithm (Model V) was used to compute the increase in the stratospheric water vapor content resulting from the oxidation of methane. The change in the vertical profile of water vapor, corresponding to increases in methane of 30 and 100 percent, respectively, is shown in Figure 7-1. These changes contribute to the indirect GWP effects of methane (IPCC). The change in the surface-troposphere radiative flux (Mode A; W/m²) due to the direct effect of methane, and that due to the considered indirect effect, as obtained by Model V, is shown below. The accompanying changes in stratospheric column ozone (ΔO₃) are also shown.

	Direct	Indirect	ΔO ₃
+30 % increase	0.18	0.04	0.7%
+100 % increase	0.52	0.12	2.1%

Indirect effects were also reported in IPCC and have also been calculated in a recent study by Lelieveld and Crutzen (1992). The amplification in the methane radiative forcing due to this indirect effect from the various calculations is summarized below. Models IV and V employ the same perturbation but differ in the radiation schemes employed:

$$RF(H_2O \text{ from } CH_4) = X \cdot RF(CH_4\text{-direct}),$$

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Model V Results

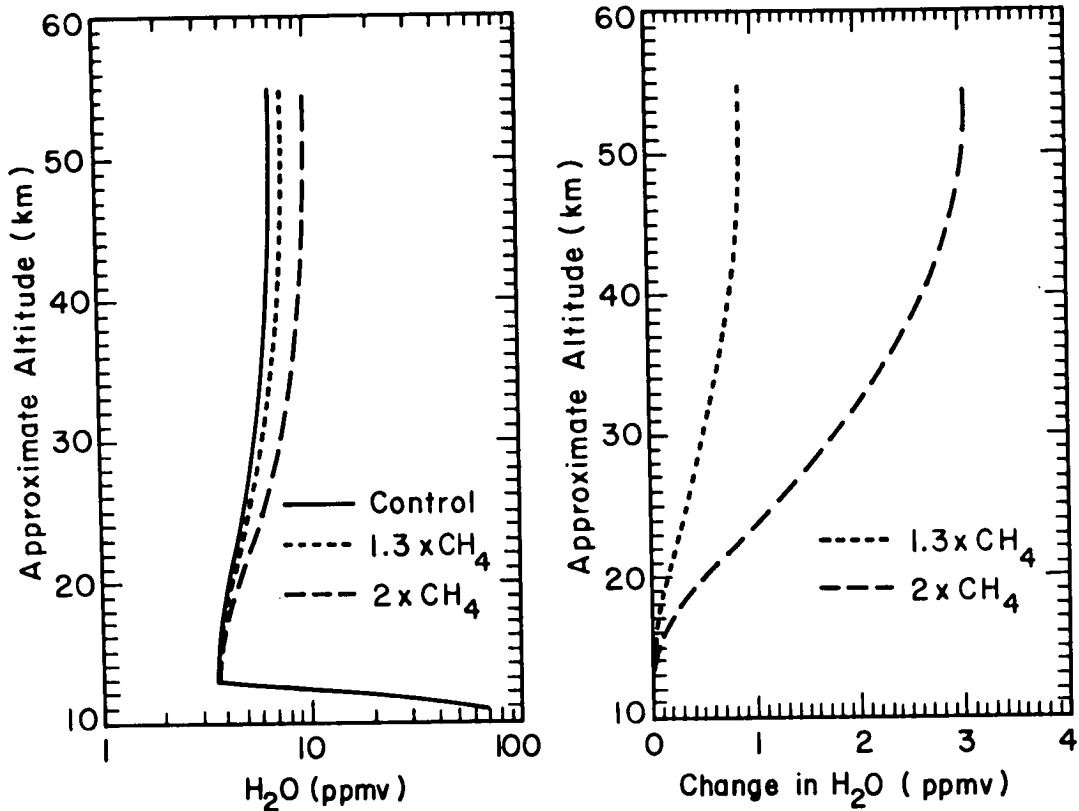


Figure 7-1 Changes in the vertical profile of water vapor due to 30 and 100 percent increase, respectively, in methane, as obtained by Model V.

where "X" reflects the enhancement in the GWP of methane and "RF" refers to radiative forcing;

- X = 0.3 (IPCC)
- = 0.22 (Model V-this report)
- = 0.31-0.38 (Model IV)
- = 0.05 (Lelieveld and Crutzen, 1992)

There are considerable differences between the results. While the first three results are approximately similar, the differences between the fourth result and the others have not been satisfactorily resolved as yet. The differences may partly reflect the different types of numerical experiments performed to calculate the effect. The difference between Model V and IV, which employed identical perturbations, are likely due to differences in the radiative transfer schemes. Further analyses are required to resolve all these differences. It is noted that the actual effect involving the oxidation of methane to water vapor will depend not only on photo-

chemical factors but also on the transport processes, and could have a latitudinal dependence.

7.3.8 Limitations and Uses of GWPs

While the GWP, as defined in IPCC, is a convenient yet reasonable practical index for ranking the relative and aggregated greenhouse gas emission impacts, it has the following limitations, some of which are very serious:

A) The radiative forcing employed in the determination of the GWP does not purport to characterize the latitudinal and seasonal dependence of the change in the surface-troposphere radiative fluxes. Nor does the GWP describe explicitly the partitioning of the surface-troposphere flux changes into that at the surface and that within the troposphere.

B) The GWP definition considers only the surface-troposphere radiative forcing rather than a particular response (e.g., surface temperature) of the climate sys-

tem. While the surface-troposphere radiative flux perturbations can be related to temperature changes at the surface in the context of the one-dimensional radiative-convective models (WMO, 1986), such a general interpretation for the temperature response in the three-dimensional general circulation models or in the actual surface-atmosphere system must be approached with caution. Further, although the GWP of a well-mixed gas can be regarded as a first-order indicator of the potential global mean temperature change due to that gas, it is inappropriate for predicting or interpreting the regional climate responses.

C) Different well-mixed gases can yield characteristically different spatial patterns of radiative forcings and climate responses (Wang *et al.*, 1991). Since the GWP is a measure of the global effect of a given greenhouse gas emission, it is most appropriate for well-mixed gases in the troposphere (*e.g.*, CO₂, CH₄, N₂O and halocarbons). The global perspective of the concept raises doubts about its applicability to species that have pronounced spatial and temporal variations such as sulfate aerosols.

D) The accounting of the indirect effects (*e.g.*, due to changes in water vapor and ozone) is more problematic than the direct effects. At present, there exist uncertainties in the details of the chemical processes as well as in the spatial and temporal variations of species involved in such transformations. Further, while the GWP concept thus far has been applied to gases with perturbations only in the long-wave spectra, it may not adequately account for the radiative effects due to inhomogeneously-distributed species with significant interactions in both the solar and the longwave spectra (*e.g.*, ozone).

E) GWP values are sensitive to significant uncertainties regarding atmospheric residence times and indirect effects. Thus, revisions to the GWP values should be expected as scientific understanding improves. In particular, the sinks for anthropogenic CO₂ emissions are poorly understood and cannot be characterized by a single exponential decay term. Because CO₂ is used as the reference gas, any revision to the calculation of its integrated radiative forcing over time will change all GWP values.

F) GWP values will change as the composition of the trace species in the atmosphere changes with time. Hence, GWP indices evaluated with respect to, say 2020, may differ significantly from those for 1990. One way of getting around this is to calculate

GWP values on the basis of an emission scenario. However, such scenarios are also subject to uncertainties so that GWP revisions due to changes in atmospheric composition cannot be avoided.

Given the above limitations, care must be exercised in applying GWPs in the policy arena. Some possible applications are identified below.

- Technology Assessment: Alternative technologies may each emit a number of greenhouse gases in different quantities. GWPs are useful in evaluating the overall greenhouse impact of these technologies. This assessment should consider the total emissions (rather than the annual emissions) affected by the technology choice.
- Emission Inventories: GWPs can be used to sum emissions of different gases to develop estimates of overall greenhouse gas emissions in "CO₂ equivalents." The different character of emissions from different sources and the different levels of uncertainty associated with the emission estimates should be considered in conducting such an analysis.
- Limitation Policies: Policies to limit greenhouse gases could be formulated in terms of the GWP of some set of gases rather than individual gases. Uncertainties in both the GWPs and the baseline emission estimates must be carefully considered in developing this approach.

GWPs should be used with caution and with other viable approaches in evaluating policy options. In particular, the use of GWPs should not substitute for scenario analyses to evaluate the implications of different emission paths over time. It is also recognized that alternative formulations of GWPs (*e.g.*, use of continuous rather than pulsed emissions) are possible and that there is necessarily no unique definition of GWPs which can satisfy all policy needs simultaneously.

7.4 RADIATIVE FORCINGS DUE TO NON-OZONE TRACE GASES (1979-1990)

Standard atmospheric profiles, including temperature and moisture (McClatchey *et al.*, 1972), are used to determine the clear-sky radiative forcing due to the increases in the non-ozone gases (CO₂, CH₄,

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N₂O and CFCs) occurring between ~1979 and 1990 (Chapter 8). The CFCs considered here include CFC-11, CFC-12, CFC-113, CFC-115, and HCFC-22. All these gases primarily interact with the terrestrial infrared (longwave) radiation. The increases in the surface-troposphere forcing (Mode A) are listed in Table 7-4.

First, it is noted that the 10 cm⁻¹ spectral resolution results from Model I have been compared with the 'benchmark' line-by-line results (Ellingson *et al.*, 1991) and found to be in excellent agreement (Ramaswamy *et al.*, 1990). Also, results from Models I and IV agree quite well with each other. The results in Table 7-4 emphasize that the trace gas radiative

Table 7-4 Clear-sky Mode A radiative forcing (W/m²) due to the increase in the non-ozone trace gases between 1979 and 1990. Both Models employ the Roberts *et al.* (1976) continuum while Model IV in addition considers the Clough *et al.* (1989) continuum (values within parenthesis). The atmospheric profiles follow McClatchey *et al.* (1972).

Case	Model I	Model IV
Tropics	0.71	0.64 (0.68)
Mid-latitude summer	0.66	0.61 (0.65)
Mid-latitude winter	0.50	0.51 (0.52)
Subarctic summer	0.60	0.57 (0.60)
Subarctic winter	0.39	0.41 (0.41)

forcing depends on latitude and season, as was pointed out for CO₂ in WMO (1986). This occurs because of the dependence of the longwave radiative transfer on the molecular absorption and emission processes, and on the Planck function, both of which, in turn, depend on the atmospheric profiles (WMO, 1986). Since no feedbacks are considered, the contrasts in the forcings should not be associated with temperature responses actually occurring at the various latitudes and during the different seasons.

An important aspect of the trace gas forcings is the vertical partitioning between the surface and the troposphere. It is known, for example that while, for CO₂, most of the effect in the surface-troposphere system is felt within the troposphere, the CFCs exert their major effect at the surface (WMO, 1986). Table 7-5 demonstrates that, in the tropics, the troposphere rather than the surface "feels" most of the instantaneous radiative forcing (Mode A) due to the trace gas

increases. This is primarily because of the large water vapor amount present in the tropical troposphere which tends to have a high optical opacity (Ramanathan *et al.*, 1979). At higher latitudes, the lesser moisture amounts serve to decrease the trapping of the flux in the troposphere and instead focus the forcing mainly at the surface. Thus, the latitudinal and seasonal dependence of the radiative forcing, and its partitioning between the surface and the troposphere, are significant aspects of the global radiative perturbations. They deserve adequate attention in the analyses of global mean quantities (*e.g.*, GWPs).

The trace gas forcings also depend on the assumptions made about the water vapor continuum absorption in the 8 to 12 micron region of the infrared spectrum. This is an important component of the longwave radiative transfer that is still not completely understood (Ellingson *et al.*, 1991). While the above discussions were with respect to one assumption

Table 7-5 Partitioning of the Mode A radiative forcing (W/m²) between surface and troposphere, as obtained by Model IV, using the Roberts *et al.* (1976) and the Clough *et al.* (1989) continuum (in parenthesis).

Case	Troposphere	Surface
Tropics	0.50 (0.48)	0.14 (0.20)
Mid-latitude summer	0.41 (0.37)	0.20 (0.27)
Mid-latitude winter	0.16 (0.09)	0.35 (0.43)
Subarctic summer	0.32 (0.27)	0.25 (0.33)
Subarctic winter	0.04 (-0.02)	0.37 (0.43)

(Roberts *et al.*, 1976), the forcing and the partitioning using another assumption (Clough *et al.*, 1989) can be different (Tables 7-4 and 7-5). This emphasizes the need for further investigations regarding the water vapor continuum. It is also pertinent to point out that there are other spectroscopic issues that have not been adequately resolved as yet which, too, can have a non-negligible bearing on the radiative forcing computations, such as assumptions about line shape and line mixing, temperature dependence of the halo-carbon absorption, etc. (see Ellingson *et al.*, 1991 for a state-of-the-art assessment of longwave radiative transfer algorithms, and IPCC, 1990).

7.5 RADIATIVE FORCING DUE TO OZONE (1979-1990)

7.5.1 Lower Stratosphere Ozone Losses (Clear Skies)

The recent Total Ozone Mapping Spectrometer (TOMS) and Stratospheric Aerosol and Gas Experiment (SAGE) results show a significant global depletion of ozone between 1979 and 1990 (Chapter 2). While the TOMS trends analysis (Stolarski *et al.*, 1991) shows a decrease in the column ozone over the middle and high latitudes of both hemispheres, ozonesonde, Umkehr and SAGE satellite analysis (McCormick *et al.*, 1992) indicate that most of this loss occurs in the lower stratosphere (Chapter 2).

In this section, the results from the trends analyses are employed to perform a series of clear-sky sensitivity studies and determine the radiative impact of the ozone decreases. These are compared with the non-ozone gas increases described in section 7-4. The study was performed for seven cases involving different geographical locations, time of the year, and atmospheric profiles (McClatchey *et al.*, 1972), as listed in Table 7-6. The conditions chosen include low and high ozone depletion cases. Two of the models listed in Table 7-1 (I and IV) performed the calculations. In both models, the ozone loss amount is confined to a layer between tropopause and 7 km above it. This is approximately consistent with the observed vertical profile of the ozone loss in the lower stratosphere.

The results under Mode A and Mode B conditions are tabulated in Table 7-7. This table lists the forcings for three different changes: CO₂+CH₄+N₂O, CFCs only and O₃ only. A histogram plot of the forcings, as obtained by Model I, appears in Figure 7-2. Results from the two models listed are in fair agreement and yield similar conclusions for Mode A. Mode B results were unavailable from Model IV.

The effects due to the non-ozone species (CO₂, CH₄, N₂O, and CFC) are solely due to their absorption bands in the infrared, with solar absorption being small and arising due to CO₂ alone. As in Table 7-4, the latitudinal dependence for the non-ozone gases is evident again. In comparing Mode B with Mode A results, an important distinction occurs for the CFCs and the other non-ozone species. Since the CFCs tend to warm the region near the tropopause (Dickinson *et*

Table 7-6 Cases selected for analyzing the radiative forcing arising due to ozone depletion in clear skies, based on the TOMS data (Stolarski *et al.*, 1991). The ozone change during the decade at the particular latitude and time, and the atmospheric profile used in the computations are listed.

Case	Latitude	Time of year	Profile*	ΔO ₃ (%)
1	0°N	July 1	T	-0.5%
2	45°N	July 1	MLS	-2.0%
3	45°N	Feb. 1	MLW	-8.0%
4	45°S	Jan. 1	MLS	-6.0%
5	45°S	Aug.1	MLW	-5.0%
6	70°N	Mar.1	SAW	-17.0%
7	75°S	Oct. 1	SAW	-32.0%

*T-Tropical, MLS-Mid-latitude Summer, MLW-Mid-latitude Winter, SAW-Subarctic Winter; see McClatchey *et al.*, (1972).

Table 7-7 Radiative forcings (W/m²) due to the non-ozone trace gas increases, CFC only increases, and that due to ozone decreases for each of the cases cited in Table 7-6. The ozone loss is assumed to occur in a layer between the tropopause and extending upward to 7 km above it. The results listed are obtained by Models I and IV (within parentheses).

Case	CO ₂ +CH ₄ +N ₂ O		CFCs only		O ₃ only	
	A	B	A	B	A	B
1	0.54 (0.54)	0.49	0.16 (0.14)	0.17	0.01 (0.02)	-0.01
2	0.51 (0.51)	0.47	0.13 (0.12)	0.15	0.08 (0.12)	-0.11
3	0.41 (0.44)	0.38	0.08 (0.09)	0.10	0.03 (0.07)	-0.42
4	0.51 (0.53)	0.47	0.13 (0.12)	0.15	0.27 (0.32)	-0.30
5	0.41 (0.44)	0.37	0.08 (0.09)	0.10	0.05 (0.07)	-0.22
6	0.34 (0.36)	0.30	0.05 (0.05)	0.06	-0.17 (-0.22)	-0.90
7	0.34 (0.36)	0.31	0.05 (0.05)	0.06	-0.41 (-0.39)	-2.38

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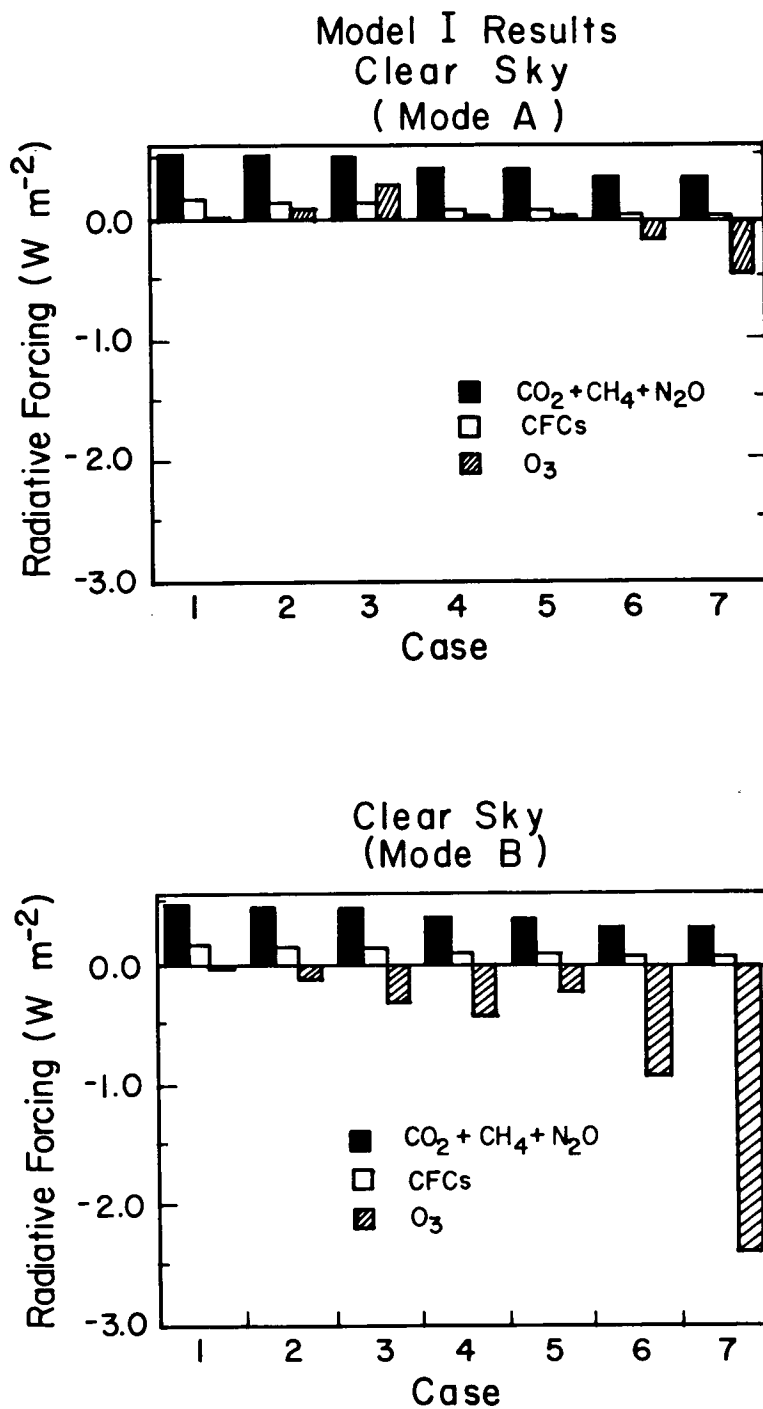


Figure 7-2 Radiative forcing due to CO₂+CH₄+N₂O, CFCs only, and that due to lower stratospheric ozone losses (see 7.5.1) at various locations and times of the year listed in Table 7-6. The seven cases employ clear-sky atmospheric profiles (McClatchey *et al.*, 1972). Upper panel shows Mode A results and the lower panel Mode B results (Model I computations). Numerical values are tabulated in Table 7-7.

al., 1978; Ramanathan *et al.*, 1985), the forcing in Mode B is associated with a warming of those altitude regimes. This leads to increased longwave emission by the lower stratosphere into the surface-troposphere system which yields a greater forcing than in Mode A. In contrast, for $\text{CO}_2 + \text{CH}_4 + \text{N}_2\text{O}$, the overall effect in the stratosphere is dominated by the cooling due to CO_2 (WMO, 1986) which consequently yields a lesser forcing in Mode B than in Mode A.

Unlike the other trace gases, changes in ozone perturb both solar and longwave radiation substantially (WMO, 1986). The effect in the solar spectrum is one of a decrease in the stratospheric absorption, resulting in more radiation becoming available for absorption in the surface-troposphere system; this constitutes a positive forcing on the system. While the solar effects due to ozone losses are determined solely by the total column ozone amounts, the longwave effects are determined both by the amount and its vertical location (Ramanathan and Dickinson, 1979; WMO, 1986; Wang *et al.*, 1980; Lacis *et al.*, 1990). In Mode A, the decrease of ozone in the lower stratosphere leads directly to a decreased emission into the surface-troposphere system which offsets, to some extent, the increase in the solar absorption there.

The reduction in the solar absorption and the change in the longwave convergence within the lower stratosphere would yield a radiatively-induced cooling tendency. This tendency due to ozone losses is similar to that due to the increases in the non-ozone gases. If this is not compensated in some manner, such as by dynamical changes, the ozone losses would lead to a cooling of the lower stratosphere (Ramanathan and Dickinson, 1979; Fels *et al.*, 1980; Lacis *et al.*, 1990). This process acts to further reduce the longwave emission into the surface-troposphere system (Ramanathan *et al.*, 1985). The solar effect is practically independent of temperature so that it is identical in both the Mode A and B determinations. However, the solar effect does affect the magnitude of the temperature change in the lower stratosphere. Thus, as far as the surface-troposphere system is concerned, both the effect of reduction in absorber amount and the reduction in temperature lead to a change in the longwave radiative transfer that competes with the solar effect. The net radiative forcing in Mode B becomes negative, being opposite in sign to the effects due to the non-ozone gases. This effect

is more marked at the higher latitudes where the ozone losses are larger. The resulting forcing at the higher latitudes acquires a substantially large negative value relative to the non-ozone forcing there. Because the forcings depend on the solar insolation, the results in both Mode A and B depend on the time of year and the geographical region.

7.5.2 Sensitivity of the Radiative Forcing to Ozone Amount and Tropospheric Cloudiness

This section highlights briefly two factors that govern the ozone radiative forcing. For this purpose, Model II was employed to analyze the quantitative influences due to the amount of ozone loss and the amount of cloudiness in the troposphere. The summary of the tests is listed in Table 7-8. The tests assume the clear-sky, mid-latitude summer (MLS) atmospheric profile (McClatchey *et al.*, 1972) and a 10 percent ozone loss as the nominal case. Solar insolation corresponds to annually-averaged conditions at the mid-latitudes. The vertical profile of the loss is as in section 7.5.1. The low, middle and high cloud amounts, their heights and properties for the 'partly cloudy' case in Table 7-8 follow the prescription in Ramaswamy and Ramanathan (1989), except that all

Table 7-8 Summary of the sensitivity tests performed using Model II to explore the dependence of the ozone-induced forcing (W/m^2). The nominal case is the clear-sky mid-latitude summer profile (McClatchey *et al.*, 1972) with a surface temperature of 294 K.

Comment	O ₃ forcing	
	Mode A	Mode B
Dependence on O₃ amount lost		
-2%	.06	-.02
-10%	.30	-.21
-20%	.61	-.48
Dependence on Clouds (10% loss)		
Clear	.30	-.21
Partly cloudy	.24	-.18
Overcast	.23	-.03

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clouds are assumed to be 'black' in the longwave. For the 'overcast case,' the high cloud amount is unity and its top is placed at the tropopause, thus yielding a condition which is the opposite extreme of a clear sky in the following sense—the upwelling longwave tropospheric flux in the lower stratosphere is the least; *i.e.*, the troposphere is optically opaque in the longwave as far as the lower stratosphere is concerned.

The results in Table 7-8 show that the net radiative forcing due to the lower stratospheric ozone losses increases markedly with the amount of ozone loss in Mode A due to the solar effect, as pointed out above. In Mode B, the temperature decreases are enhanced with increases in the ozone loss, thus yielding increasingly negative ozone forcings. The experiments with clouds indicate that Mode B results are sensitive to cloud properties. This is due to the upwelling flux from the troposphere playing an important role in the changes in the longwave flux convergence within the lower stratosphere. This, in turn, determines the magnitude of the temperature reduction there, with a concomitant effect upon the changes in the stratospheric longwave emission into the troposphere. The larger the upwelling flux from the troposphere, the larger are the resultant effects. Thus, for the extreme overcast case, there results the least forcing, and the difference from the clear-sky results is quite pronounced. Although the studies here are by no means exhaustive, they indicate that the Mode B results, owing to their dependence on the lower stratospheric temperatures, are sensitive to tropospheric radiative influences. In contrast, since Mode A does not depend on the temperature feedbacks in the stratosphere, it is not influenced by the tropospheric longwave processes.

7.5.3 Lower Stratospheric Ozone Losses (General Atmospheric Conditions)

As the third part of the stratospheric ozone loss study, the sensitivity examination was extended to study the zonal impacts due to the decadal ozone decreases for general atmospheric conditions. Also considered again for comparisons are the forcings due to greenhouse gases other than ozone. Cloud amounts in the troposphere for Models I and III follow London (1957) while the Stowe *et al.* (1989) climatology is adopted in Model IV. All the models employed in the calculations of this section have dif-

ferent initial climatologies of ozone and water vapor. The altitude range of ozone loss employed at each latitude for the model simulations are again similar to those employed in section 7.5.1. The latitudinal dependence of the column ozone loss amounts follow the TOMS percentage trends at each latitude. The vertical distribution of the loss applied differs among the models. Models III and IV applied equal ozone loss amounts in each layer within the altitude of depletion while Model I applies an equal percentage loss in the concerned layers. Also, Model I employs an ozone climatology that, when combined with the TOMS percentage trends, may be overestimating the column loss amount at latitudes poleward of 70 degrees. Most of the results in this section pertain to January and July conditions. However, Models I and IV have simulated the perturbations corresponding to all four seasons and these are considered below in some of the discussions. (Note: A slightly different sensitivity experiment, which leads to the same implications as discussed below, is reported in Ramaswamy *et al.*, 1992).

The Mode A results, as obtained by Models I and IV, are shown, respectively, in Figures 7-3 and 7-4. Note that Figure 7-3 illustrates the non-ozone and ozone forcings separately while Figure 7-4 illustrates the non-ozone and the non-ozone + ozone forcings. The characteristics of Mode A, as in section 7.5.1, can be summarized as follows: there is an increase of solar radiation into the surface-troposphere system, and there is a decrease of the longwave emission, with the net effect dependent on the latitude. The impacts on the surface-troposphere forcing are greater for latitudes having larger ozone losses.

The Mode B results have been obtained by Models I and III (latter only for Northern Hemisphere) and the results appear in Figures 7-5 and 7-6, respectively. Note that Mode B results were unavailable from Model IV while Mode A results from Model II are not plotted. The forcings are shown for the non-ozone gases taken together, and for the ozone losses. Model III yields somewhat more negative ozone forcings than does Model I poleward of 30° in the Northern Hemisphere. Both models indicate a heightened sensitivity with latitude. The most important reason for this feature is the increasing ozone loss at the higher latitudes. The differences between Models I and IV in Mode A and between Models I and III in Mode B are attributable in general to a

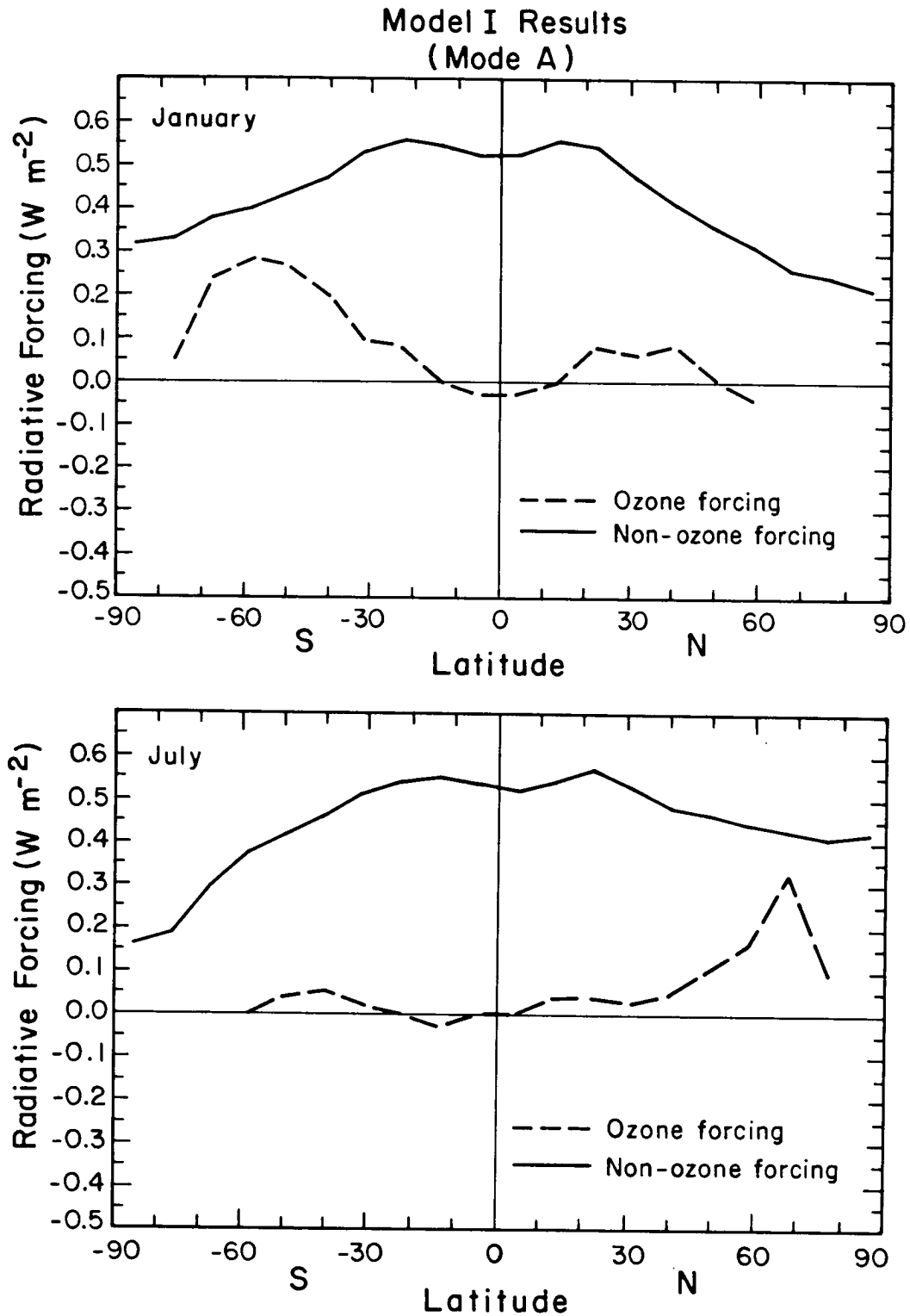


Figure 7-3 Mode A January and July radiative forcing as obtained by Model I for the 1979 to 1990 increases in all the non-ozone gases (Chapter 8) and that due to the lower stratospheric ozone losses (Stolarski *et al.*, 1991; McCormick *et al.*, 1992).

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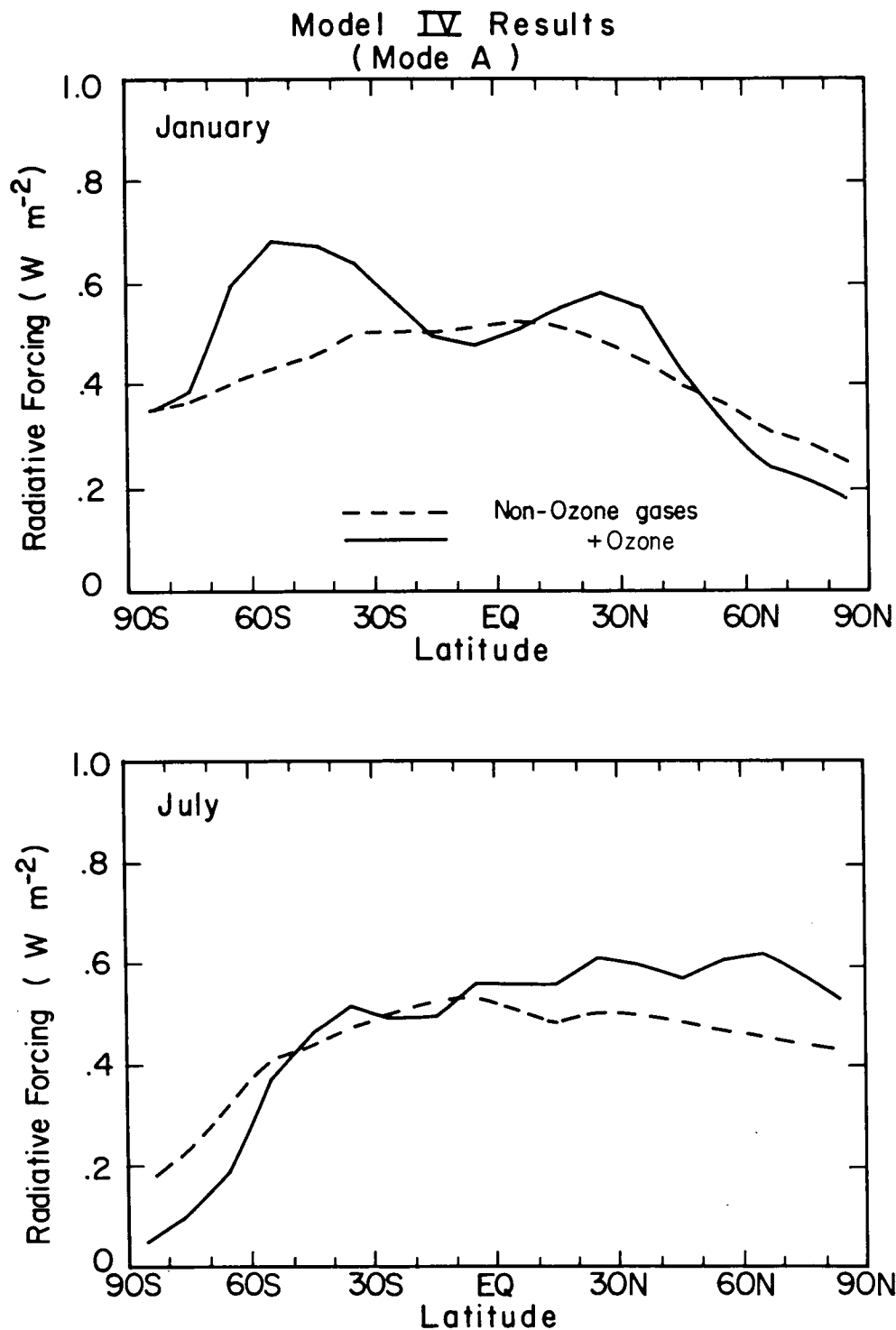


Figure 7-4 Same as Figure 7-3, except as obtained by Model IV. Note that the results are plotted for non-ozone gas increases only, and for non-ozone + ozone changes.

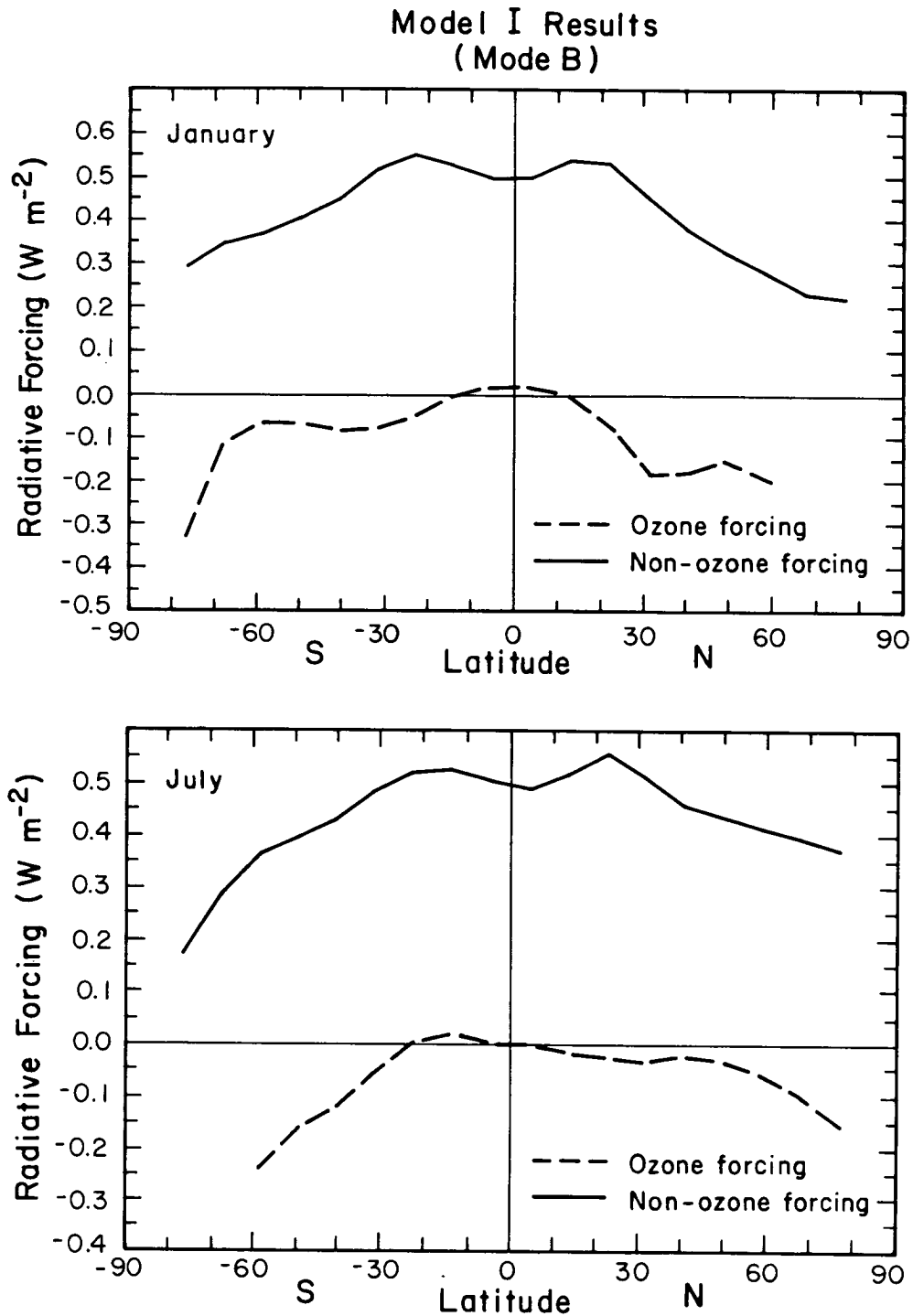


Figure 7-5 Mode B January and July radiative forcings due to non-ozone gas increases and ozone losses as obtained by Model I. Compare with the corresponding Mode A results in Figure 7-3.

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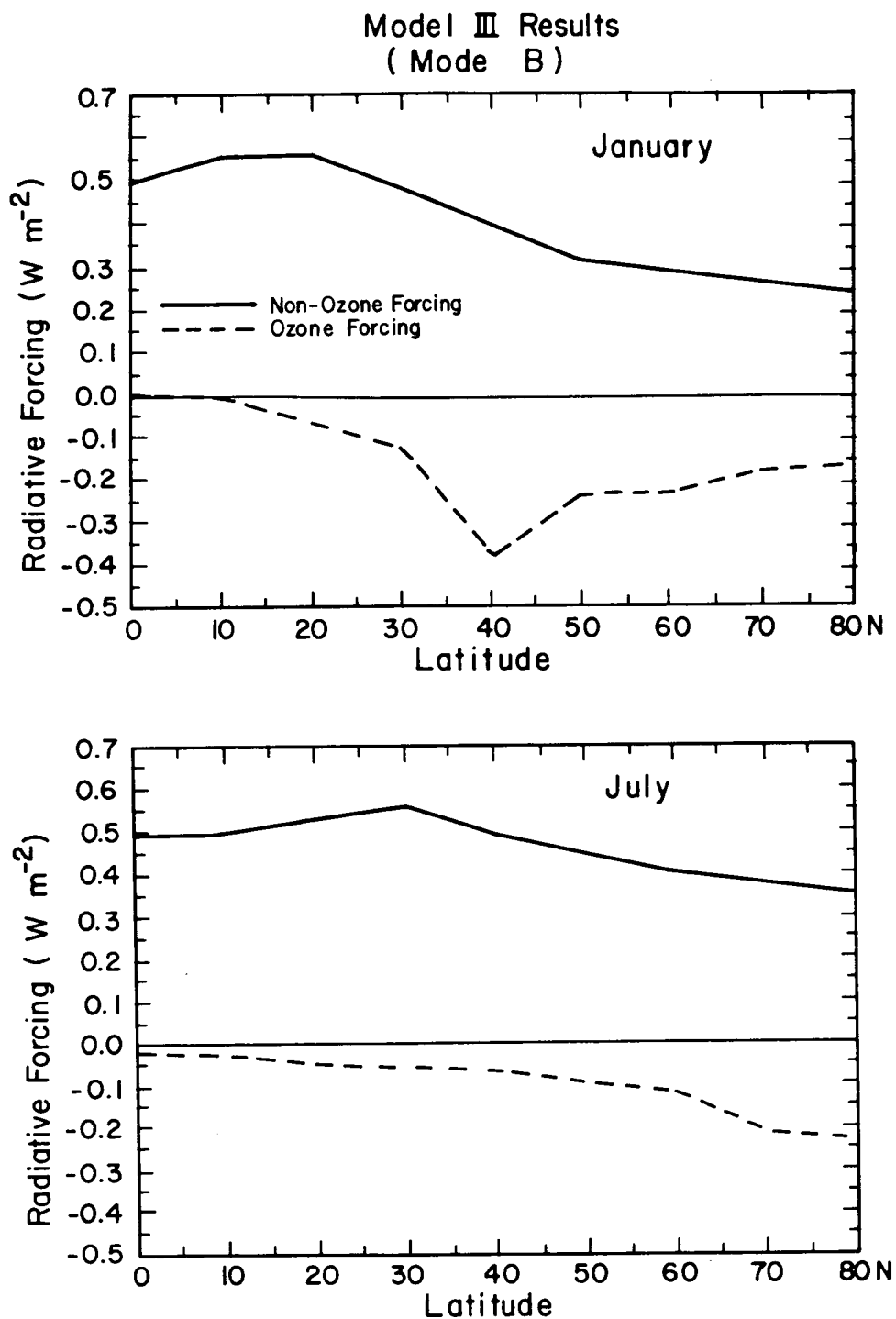


Figure 7-6 Mode B January and July radiative forcings due to non-ozone gas increases and ozone losses as obtained by Model III.

number of factors, *e.g.*, different initial climatologies including tropospheric conditions, vertical resolution, tropopause location, vertical distribution of the ozone loss, radiative transfer algorithms, etc.

The temperature change in the lower stratosphere at $\sim 40^\circ\text{N}$ (January), obtained in Mode B after stratospheric equilibrium in the presence of a fixed dynamical heating, is shown in Figure 7-7. The temperature decrease due to the ozone losses in the lower stratosphere exceeds substantially that due to the non-ozone gas increases. Comparing Models I and III, the temperature decrease in the latter is shifted more to the lower altitudes in the stratosphere because of the manner of the vertical distribution of the loss, as mentioned above. Specifically, Model III has a greater amount of loss in the lower portion of the stratosphere than does Model I. It is this ozone-induced temperature decrease that is responsible for the enhancement in the decrease of the stratospheric longwave emission to the troposphere in Mode B, thus resulting in a net negative surface-troposphere radiative forcing at this latitude.

To emphasize the relative importance of the 1979 to 1990 Mode B ozone radiative forcings, the ratio of the ozone to all the non-ozone forcing and that of ozone to CFC only forcing, as obtained by Model I, is shown in Figure 7-8. Poleward of 30 degrees, the magnitude of the (negative) ozone forcing become increasingly comparable to and can even exceed the (positive) CFC forcing. At these latitudes, the ozone forcings can also be a significant fraction of the (positive) non-ozone forcing during all seasons and in both hemispheres.

The globally averaged forcings due to changes in the radiatively active gaseous species between 1979 and 1990 are listed in Table 7-9. They suggest that the instantaneous ozone forcing, without any stratospheric temperature feedback (Mode A), enhances the direct CFC radiative forcing by more than 40 percent. However, considering the stratosphere to be in equilibrium, which is more appropriate to derive the climatically significant surface-troposphere forcing (WMO, 1986; IPCC), and assuming that there is no change in the dynamical heating of the stratosphere. (Mode B), the ozone forcing is comparable but opposite in magnitude to the CFC forcing, besides being about 20 percent of the entire 1979 to 1990 trace gas forcing.

7.5.4 Characteristic Features of the Ozone Forcing

The forcing due to ozone is unique in two respects when compared to the forcings by the other gases (section 7.4). First, although there is a global mean offset of the CFC direct forcing by the ozone losses for the 1979 to 1990 period (Table 7-9, Mode B results), it is evident from Figures 7-5, 7-6 and 7-8 that this arises because of a significant negative forcing by ozone occurring in the higher latitudes only. In particular, the radiative forcing due to CFCs + ozone changes, as inferred from Figure 7-8, ranges from a net positive one at the low latitudes to a net negative one at the higher latitudes for the period considered. Further, the ozone losses suggest that the latitudinal dependence of the forcing due to changes in all trace gases between 1979 and 1990 could have had a meridional gradient quite different from that expected for the non-ozone species only.

Table 7-9 Globally and annually averaged radiative forcing (W/m^2) of the surface-troposphere system due to changes in the concentrations of the trace gases between 1979 and 1990: a) increases in CFCs only, b) combined effects of increases in CO_2 , CH_4 , N_2O and CFCs and c) decreases in ozone. Results obtained from both Mode A and B are listed. Model I averages employ each of the four midseason months in both hemispheres. Model III results are only for January and July perturbations in the Northern Hemisphere. Model IV results are available only for Mode A averaged over the four seasons.

Mode A			
Model	CFCs	All non-ozone gases	Ozone
I	0.09	0.49	0.06
III	0.10	0.51	0
IV	0.07	0.51	0.03
Mode B			
Model	CFCs	All non-ozone gases	Ozone
I	0.10	0.46	-0.08
III	0.11	0.47	-0.11

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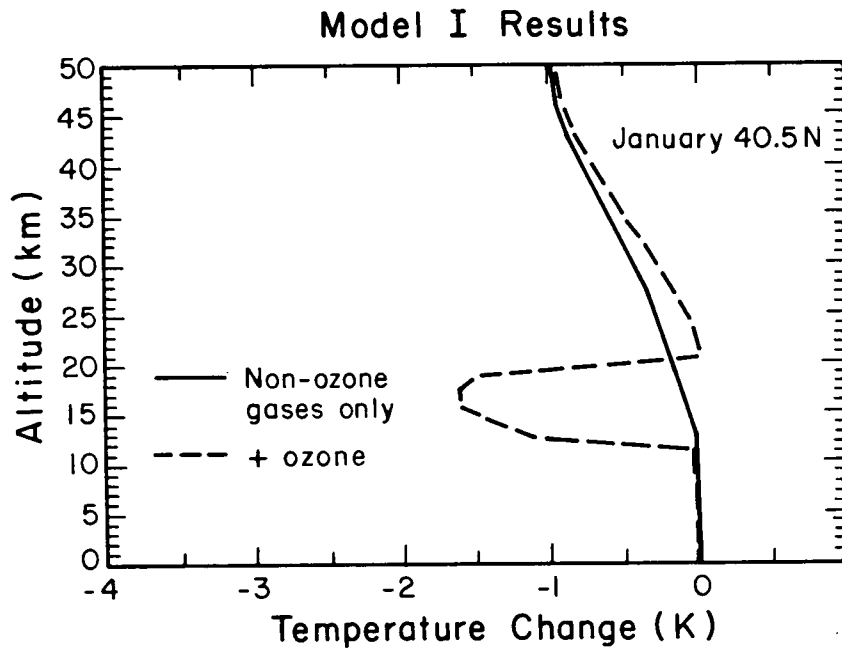
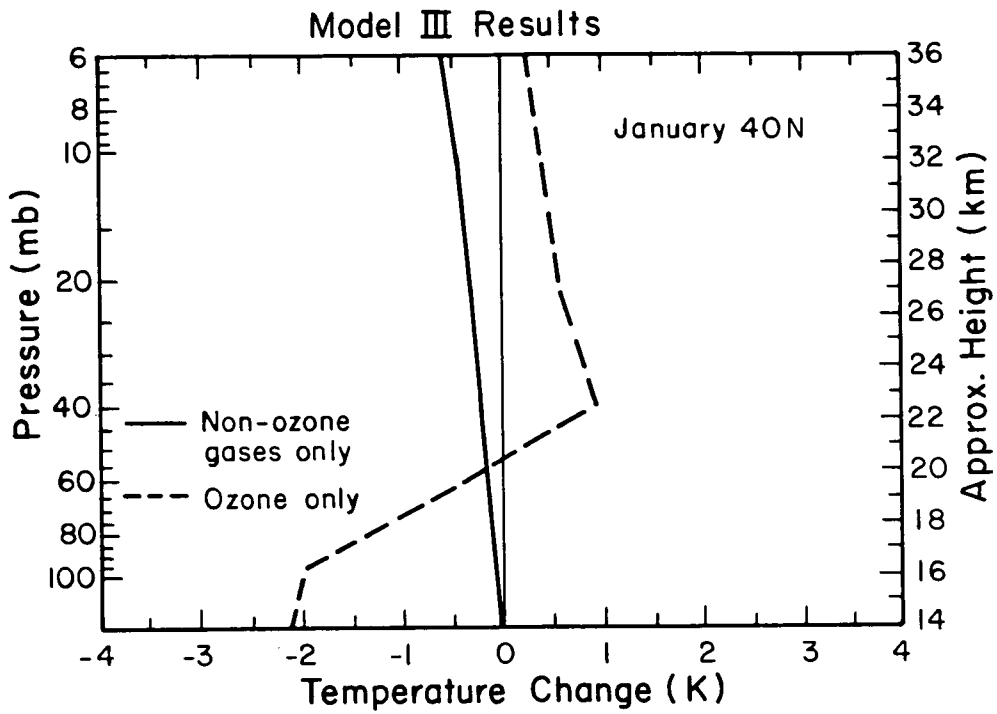


Figure 7-7 Temperature change at ~40°N for January conditions as obtained by Model III and Model I due to increases in the non-ozone gases (Chapter 8) and due to the lower stratospheric ozone losses occurring between 1979 and 1990 (Stolarski *et al.*, 1991; McCormick *et al.*, 1992). Note that, for Model III, the results are plotted for the two components separately while, for Model I, the results are plotted for non-ozone gases only, and for ozone + non-ozone gases.

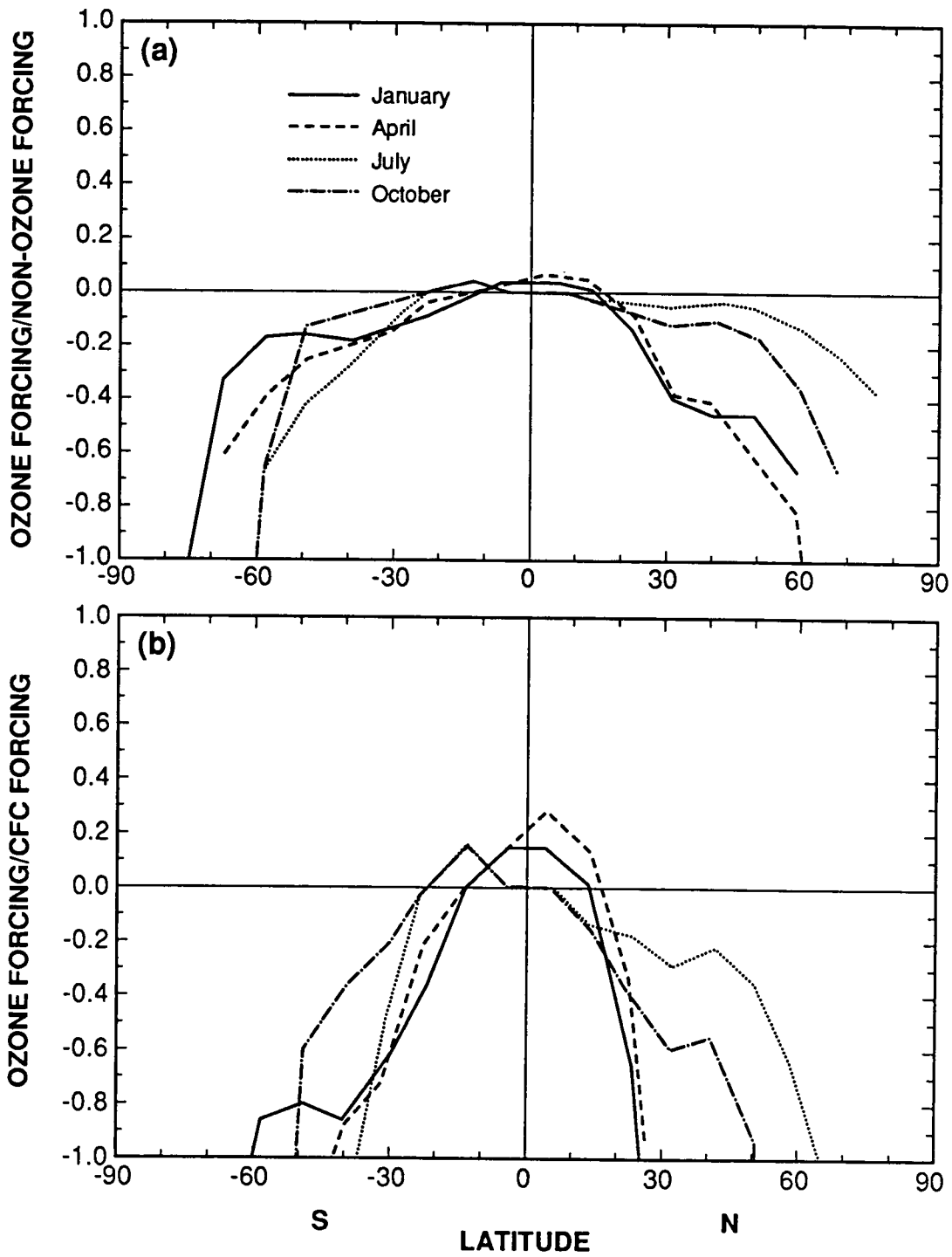


Figure 7-8 Ratio of the Mode B ozone forcing to that due to a) the non-ozone gas increases and b) CFCs alone for all seasons, as obtained by Model I.

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There is yet another unique aspect concerning the ozone forcing which concerns the partitioning of the effect between the surface and the troposphere. For ozone, unlike the other radiatively active species, both solar and the longwave interactions are significant. As shown by Ramanathan and Dickinson, (1979), most of the solar forcing is "felt" at the surface, while the longwave forcing is "felt" primarily within the upper troposphere. While that study considered a different ozone loss profile, yet the same features are seen even for the present lower stratospheric losses. Table 7-10 indicates this to be true in both Mode A and Mode B. The negative Mode B surface-troposphere forcing at the mid-to-high latitudes thus consists of a dipole-like feature, with a solar-induced warming at the surface, combined with a longwave-induced cooling tendency of the troposphere. There is a greater decrease for the troposphere in Mode B than in Mode A due to the temperature decreases in the lower stratosphere. Comparing with the effects for the non-ozone gases (Table 7-5), ozone stands out not only in terms of being able to cause a negative radiative forcing (Lacis *et al.*, 1990),

but also in imparting a distinctively different vertical partitioning of this forcing.

The negative surface-troposphere Mode B forcing due to ozone can be interpreted as a cooling of the surface only if the convective coupling between the surface and the troposphere is strong, as is assumed in one-dimensional radiative-convective models. In the polar regions and for the upper troposphere, this need not be true (Dickinson *et al.*, 1978; Ramanathan and Dickinson, 1979; WMO, 1986). Three-dimensional General Circulation Model (GCM) studies are required to comprehensively determine the effect of this coupling on the response of the climate system to the ozone forcing.

7.5.5 Greenhouse Implications of the 1979 to 1990 Observed Ozone Losses

Since the ozone losses are suspected to be due to heterogeneous chemical reactions involving chlorine- and bromine-containing chemicals, the Mode B results here suggest that the ozone-depleting substances, which include the anthropogenic emissions of CFCs, have substantially reduced the radiative contributions of the CFCs to the greenhouse forcing over the past decade. Thus, the overall greenhouse effect attributed to the CFCs taken together must recognize this potential indirect contribution due to the chemically induced destruction of ozone.

The Mode B results here for ozone are in the same sense as those estimated by Lacis *et al.* (1990) for mid-latitudes during the decade of the 1970s. They are, however, different from those obtained using one-dimensional gas-phase only photochemistry models that predicted losses of ozone primarily in the middle and upper stratosphere (Ramanathan *et al.*, 1985) and led to a positive forcing. Since the observed ozone losses over the 1979 to 1990 period are indicating losses in a different region of the stratosphere than predicted earlier, this leads to a substantially different ozone radiative impact, and a different implication about the overall effect of the CFCs taken together in the global greenhouse forcing. The differences are entirely due to the differences in the assumed vertical profile of the ozone loss.

It is emphasized that the ozone forcing are extremely sensitive to the altitude of the losses (Ramanathan *et al.*, 1985; Lacis *et al.*, 1990). In the

Table 7-10 Mode A and Mode B surface and troposphere ozone forcing (W/m^2) in different latitude belts, as obtained by Model I.

Latitude	Mode A	
	Surface	Troposphere
90-60°S	0.31	-0.13
60-30°S	0.16	-0.05
30-10°S	0.006	0.
10S-10°N	-0.02	0.
10-30°N	0.04	0.
30-60°N	0.15	-0.06
60-90°N	0.25	-0.10

Latitude	Mode B	
	Surface	Troposphere
90-60°S	0.27	-0.76
60-30°S	0.15	-0.24
30-10°S	0.005	-0.01
10S-10°N	-0.02	-0.03
10-30°N	0.04	-0.07
30-60°N	0.14	-0.26
60-90°N	0.22	-0.49

computations here, it has been assumed that all the losses occur in the lower stratosphere. Detailed analyses from SAGE (McCormick *et al.*, 1992) and ozonesondes (Chapter 2) reveal that there are additional losses occurring in the middle stratosphere (~40 km) together with a small increase at ~30 km. There still is some uncertainty regarding the exact profile and the magnitude of the loss in the immediate vicinity of the tropopause. The SAGE profiles are available only from ~17 km and above globally, and indicate an increasing percentage of loss with decreasing altitude in the lower stratosphere. While the assumptions in the model sensitivity calculations of this section serve to infer the general implications of the radiative forcing due to lower stratospheric ozone losses, more precise estimates of the climate forcing need to consider carefully the details of the vertical loss profile. As an example of this sensitivity, if the hypothetical assumption were made that the TOMS observed losses are uniformly distributed in the entire stratospheric column (as in WMO, 1986), then the longwave effects become considerably less than the one obtained in the Mode B calculations here, leading to a small global ozone forcing (-0.01 W/m² from Model I and -0.04 W/m² from Model III; compare with Table 7-9). Thus, inferences about ozone forcings depend crucially on both the total column change as well as the change in the vertical profile and both these entities need to be monitored very carefully.

7.5.6 Effect of Dynamics on Stratospheric Temperature Changes

The indirect effect of CFCs on the climate system due to depletion of ozone in the lower stratosphere is critically sensitive to the actual temperature change and its distribution in the lower stratosphere. While Mode B generally provides more realistic assessments of forcing than Mode A, it too, may be unrealistic if the predicted stratospheric temperature change is not realized. Because atmospheric circulation can change in response to radiative perturbations, the dynamical contribution to the heating could also change, thereby contributing to the actual temperature change (Dickinson, 1974). This dynamical contribution is most likely to be significant when the forcing has strong spatial gradients, as does the observed ozone depletion.

A better estimate of temperature change may be obtained from three-dimensional general circulation models combined with careful diagnostic studies of the behavior of the real atmosphere. The observed global ozone depletion has not yet been simulated in a GCM, but simulations for the following scenarios of ozone changes have been performed: a) a uniform decrease of O₃ throughout the stratospheric column (Fels *et al.*, 1980; Kiehl and Boville, 1988), b) a homogeneous gas phase chemical model prediction of ozone depletion (Kiehl and Boville, 1988), which is different from the observed losses, and c) observed springtime depletion in the Antarctic region (Kiehl *et al.*, 1988). The resulting stratospheric temperature changes in these studies indicate that, unless the column depletions are large (>50 percent), the fixed dynamical heating (FDH) temperature response (*i.e.*, Mode B) closely resembles the GCM response, but there are some latitude-dependent departures. The GCM studies of Rind *et al.* (1990, 1991) show that dynamically forced temperature changes can result from subtle interactions between changes in the atmospheric structure, upper tropospheric latent heat release, and the forcing and transmission of planetary waves and gravity waves.

Lacking appropriate GCM simulations corresponding to the observed ozone loss, we turn to the observed temperature trend for an assessment of the stratospheric temperature changes due to the observed ozone depletion. Unfortunately, this trend is ambiguous because of the large natural variability in the lower stratospheric temperatures. Two different analyses suggest that there has been a cooling of the global lower stratosphere (50–100 mb layer) over the past two decades, which is more significant in the southern polar regions and less so for the Northern Hemisphere mid-latitudes. For the 1973–1987 period, Angell (1988) obtains a global mean trend of -0.62 ± 0.48 K/decade while, for the 1964–1988 period, Oort and Liu (1992) obtain a global mean trend of -0.40 ± 0.12 K/decade. The global mean of the decadal temperature change, averaged over the 50–100 mb layer, as obtained in the Mode B results of section 7.5.3, is -0.46 K from Model I and -0.63 K from Model III. Although the model results appear to be within the uncertainty limits present in the global mean trends, it must be recognized that there are other physical factors (*e.g.*, changes in tropospheric state, aerosols, etc.), not accounted for here, which could also be

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contributing to the trends. Thus, the question of the horizontal and vertical distribution, and the magnitude of the stratospheric temperature change at each latitude produced by the observed ozone depletion remains open and requires further investigations, including GCM simulations.

7.5.7 Sensitivity to Increases in Tropospheric Ozone

Thus far, only the stratospheric decrease in O_3 has been considered. However, there is some evidence from ground-based stations of increases in tropospheric ozone during the past decade (Chapter 2). The greenhouse effect of tropospheric ozone has been anticipated and investigated in earlier reports (WMO, 1986). Even though tropospheric ozone amounts are less than the stratospheric ones, the effective long-wave optical depths are greater in the troposphere (Ramanathan and Dickinson, 1979).

In order to investigate the forcing upon the surface-troposphere system due to increases in tropospheric ozone, Model I performed a sensitivity study which employs the stratospheric loss at $40^\circ N$ as a base. Tropospheric ozone is then increased at this latitude from 0 to 17 percent, the latter being the reported decadal increase by the Hohenpeissenberg station. The vertical profile of the tropospheric ozone change follows the Hohenpeissenberg trends. Figure 7-9 illustrates the reduction of the stratospheric impacts due to increases in tropospheric ozone. There is almost a linear increase of the greenhouse effect with a percentage increase in the tropospheric ozone concentration. Although the greenhouse effect of increases in tropospheric ozone can be significant, there is at present insufficient evidence that such changes are taking place globally, especially in the radiatively important upper tropospheric regions (Lacis, *et al.*, 1990). In the absence of statistically significant global trends, it is not possible at this stage to quantify the global greenhouse effect due to tropospheric ozone.

7.6 RADIATIVE FORCING DUE TO TROPOSPHERIC SULFATE AEROSOLS

Sulfur-containing gases emitted into the atmosphere through natural and anthropogenic processes do not contribute significantly to the greenhouse effect. However, a large fraction of these gases are

transformed into sulfate aerosol particles in the atmosphere. These aerosol particles may affect climate in several ways, the most important being:

- scattering of sunlight back to space in clear skies, thereby reducing the amount reaching the surface,
- increasing the number of cloud condensation nuclei (CCN), thereby potentially altering the physical characteristics of clouds (*e.g.*, albedo, precipitation), and
- altering the chemical balance (*e.g.*, ozone concentrations) by providing surfaces for chemical transformations. This could be particularly important in the lower stratosphere.

The most important new information that has become available since IPCC refers to the first of the above three mechanisms. Based on simulations of the global distribution of sulfate aerosols, Charlson *et al.* (1990, 1991) used previously acquired information on backscattering coefficients per unit mass of sulfate to estimate the impact on the shortwave radiation balance of anthropogenic sulfur emissions. They concluded that the clear-sky effect alone, averaged over the Northern Hemisphere, corresponds to a negative forcing at the Earth's surface of approximately 1 W/m^2 . This is comparable to and of the opposite sign to the forcing due to the buildup of CO_2 in the atmosphere. Even though this estimate needs to be refined through further calculations and measurements, it is based on relatively few basic assumptions and is probably not uncertain by more than a factor of 2.

A very important implication of this finding is that the effective anthropogenic climate forcing in the Northern Hemisphere during the past century is likely to have been smaller than previously believed. A meaningful quantitative comparison between the positive forcing of the well-mixed greenhouse gases and the negative forcing due to sulfate aerosols is complicated by the nonuniform spatial distribution of the latter, in contrast to the uniform geographical distribution of the well-mixed gases.

Anthropogenic sulfur emissions in the Southern Hemisphere are much smaller and the resulting concentrations correspondingly lower so that no compensating cooling tendency is likely to have taken place there. The expected hemispheric difference in forcing is qualitatively consistent with an observed small difference in temperature trends between the hemi-

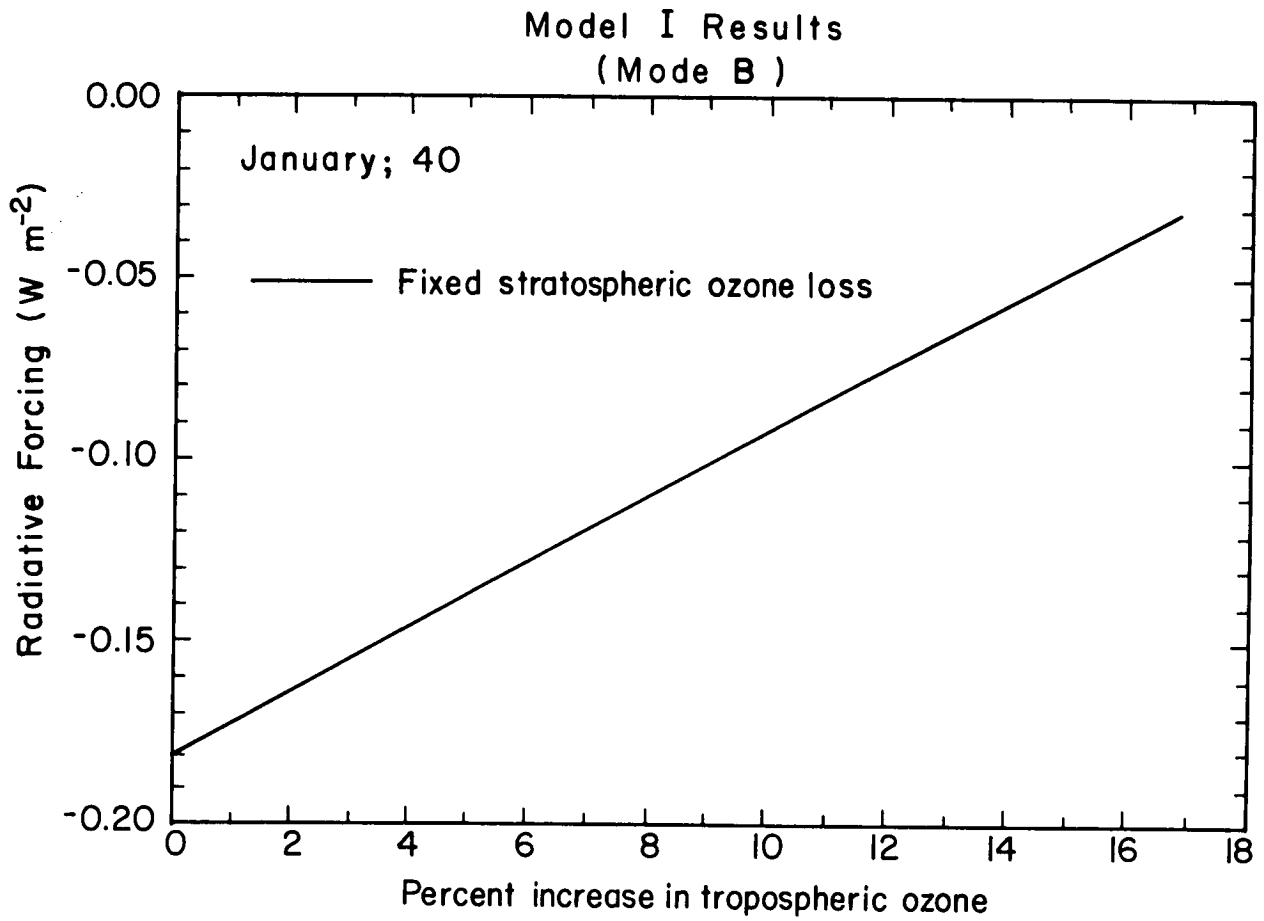


Figure 7-9 Radiative forcing due to increases in tropospheric ozone as obtained from Model I. The vertical profile of tropospheric ozone changes conforms to the Hohenpeissenberg observations. The base state (*i.e.*, the forcing for a 0 percent increase) corresponds to the stratospheric losses at 40°N for January conditions (see section 7.5.7).

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spheres over the past century (Wigley, 1989). A more detailed analysis will also have to take into account the potential effect of aerosol particles from biomass burning in both hemispheres (Crutzen and Andreae, 1990).

Further changes in the forcing due to sulfate aerosols and greenhouse gases will depend on the variations in their respective emission sources. Because of the short atmospheric residence times of sulfates and their precursors, the atmospheric concentrations will adjust within weeks to changes in emissions. This is different from the case for the greenhouse gases which have effective lifetimes ranging from decades up to centuries. For example, the concentration of CO₂ will continue to rise for more than a century even if emissions are kept constant at today's level. This difference has been examined by Wigley (1989) and Charlson *et al.* (1991), who show how the climate forcings due to CO₂ and sulfate aerosol would change if the global fossil fuel consumption leveled off and was eventually reduced. Because of the rapid growth in emissions during the past decades, both the greenhouse forcing due to CO₂ and the opposite forcing due to aerosols have grown accordingly. During a leveling off phase, the greenhouse forcing will continue to grow whereas the aerosol forcing will remain constant. During a decay phase, the greenhouse forcing will start to level off and the aerosol forcing will decline. This simple example demonstrates that the relative importance of these two major anthropogenic forcing agents in the future will depend critically on changes in the use of fossil fuel (large-scale desulfurization measures would, of course, also have to be considered).

Because of the very different character of the forcing due to aerosols as compared to that of the well-mixed greenhouse gases, no attempt is made to define a GWP for anthropogenic sulfur emissions. It is also considered that there is no meaningful use of a GWP value for sulfur; trade-offs between reduction in greenhouse gases and increases in sulfur emissions do not seem reasonable.

7.7 RADIATIVE FORCING DUE TO STRATOSPHERIC AEROSOLS

Observations of the past decade (lidar, satellite, balloon, sunphotometer) indicate that the stratospheric aerosol concentration remained higher than that measured in 1979 (a relatively quiescent period) throughout

most of the 1980s. This was probably in part attributable to the major El Chichón volcanic eruption in 1982 plus few other minor ones (McCormick and Trepte, 1987). An additional contribution could also have been due to anthropogenic means (Hofmann, 1990). With the recent major eruption of the Mt. Pinatubo volcano, there is now a fresh accumulation of particulates in the stratosphere that can be expected to yield a radiative forcing (WMO, 1990). The radiative effects may be already manifest in the tropical lower stratosphere where there has been a temperature increase of 3–4 K (McCormick, M. P., private communication) in the 2 months following the eruption. These aerosols can also be expected to exert a radiative forcing on the surface-troposphere system. It has been suggested that global coolings have occurred following major eruptions in the past but are of the order of a few tenths of a Kelvin or less (Mass and Portman, 1989). In the case of the aerosols from the Mt. Pinatubo eruption, the magnitude of the surface cooling, if any, has yet to be confirmed.

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