Final Report on

**Assessing the Climatic Effect of
Carbon Dioxide and Other Trace Gases Using an
Interactive Two-dimensional Climate-Chemistry Model**

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Appendix E: MacKay, R.M., M.K.W. Ko, R-L Shia, S. Zhou, and Y. Yang (1996) The response of a coupled climate-chemistry model to greenhouse forcing. Manuscript in preparation.

Appendix F: MacKay R.M. and M.K.W. Ko (1996) Normal Modes and the transient response of the climate system. Manuscript in preparation.

Appendix G: Ko, M.K.W., R.M. MacKay, R-L Shia and N-D Sze (1996) An Estimate of global warming from chlorofluorocarbons and their alternatives over the next century. manuscript in preparation. The global change of temperature and ozone simulated by a coupled climate-chemistry model.

ABSTRACT

Carbon dioxide (CO₂), water vapor (H₂O), nitrous oxide (N₂O), methane (CH₄), and chlorofluorocarbons (CFCs) and their substitutes have been identified as greenhouse gases. Changes in the concentrations of these long-lived gases in the atmosphere are expected to affect the earth's climate. Increases in the concentrations of CFCs, N₂O, and CH₄ can also perturb chemical cycles and modify the distribution of ozone, which is itself a greenhouse gas. There is also the possibility that surface warming resulting from increases of CO₂ and other trace gases could modify the transport circulation in the lower stratosphere, leading to ozone redistribution. In the recent IPCC report, the role of tropospheric aerosols, stratospheric aerosols, and natural solar variability have also been identified as having sizable effects on climate, both by direct perturbation of the radiative balance and indirectly by changing ozone. Although the effect of changing CO₂ is by far the dominant factor on a century time scale, the effects from the other identified factors are important on a decade time scale. It is important to understand the mechanisms that relate these changes to climatic responses. Developing appropriate numerical models with the capability to simulate these mechanisms will enable us to correctly interpret the observed climate changes that have occurred to date, as well as predict future changes in climate. It is presently impractical to run comprehensive 3-D general circulation model simulations of the interactions between atmospheric chemistry and the rest of the climate system on time scales of decades to centuries. Thus, 2-D models and other lower resolution models play an essential role in understanding the complex interactions of the integrated climate system.

We have addressed two issues using a suite of tools. For the problem of chemistry-climate coupling, where the underlying physics of the problem requires that studies be performed with multi-dimensional models, we developed a two-dimensional zonal-mean interactive climate-chemistry (ICC) model with a latitudinal resolution of 10° latitude. The chemistry module is based on the AER interactive chemistry-dynamics (ACD) model of Schneider et al. (1989) and Ko et al. (1993). The climate module is based on the AER three-zone seasonal-radiative-dynamical (ASRD) model of Wang et al. (1990) and Molnar et al. (1994). In the previous 8 months, we concentrated on analyzing model results to find out how the different components combine to define the feedback sensitivity of the model to changes in radiative forcing. To help us focus on impacts of the individual greenhouse gases on a decadal time scale, we use a simple multi-box ocean model to derive the normal mode response of the atmosphere and ocean system to applied radiative forcing. Using this model, we derive the expected contribution of each greenhouse gas to global warming in the next several decades.

Our study supports the idea that further refinement of the 2-D climate-chemistry model will allow us to examine the total effect of solar variation and stratospheric aerosols, including the indirect effect from changes in ozone. The flexibility of the 2-D model allows us to perform realistic scenario calculations corresponding to historical emission rates as well as different control strategies proposed by UNEP, IPCC, and other national and international agencies. The results will ultimately lead to a better understanding of the validation of model predictions and detection of signals for greenhouse warming that might have already occurred. Our work will also provide useful information for the design of reduced-form models that will be used in integrated studies.

1. SCIENTIFIC RESULTS

1.1 Scientific Background

Current projections indicate that by the middle of the next century the earth's climate may be considerably different from that of today, due to increasing concentrations of greenhouse gases and accompanying changes in aerosols and clouds (see e.g., IPCC, 1994). Although an increase in atmospheric concentration of CO₂ is the major drive for the predicted greenhouse warming over the century time scale, changes in the concentrations of other greenhouse gases (H₂O, CH₄, N₂O, O₃, CFCs as well as their substitutes) are expected to augment or ameliorate the CO₂-warming significantly on a decadal time scale. Changes in radiative forcing due to changes in CFCs in the past two decades are comparable to that caused by changes in CO₂ in the same period. Thus, understanding their relative contribution is particularly important for interpretation of the observed climate change in the past decades.

The radiative budget of the earth-atmosphere system involves the balance of the net incoming solar radiation and the outgoing longwave radiation at the top of the atmosphere. We are relatively confident about the magnitudes of these quantities because of the availability of satellite measurements. The energy balance at the earth's surface is more complicated, however. The amount of solar radiation reaching the ground is modulated by scattering and absorption by trace gases, aerosols, and clouds. The amount of solar energy absorbed is modulated according to the ground albedo. Most of the longwave radiation emitted by the surface is absorbed in the atmosphere and re-radiated. Energy deposited into the atmosphere via latent heat exchange is also transformed to longwave radiation. The radiative balance differs from latitude to latitude because of differences in solar zenith angle, land-ocean distributions, albedo, cloud cover, and in the amount of trace gases and aerosols. The atmosphere cannot be easily treated as a number of independent column atmospheres because of the transport of heat within the atmosphere and in the ocean. However, the transport effect can be parameterized in two-dimensional (latitude-height) models.

Previous studies of the greenhouse effect have concentrated on the temperature response to CO₂ loading in the belief that the responses for other greenhouse gases are proportional to the respective change in radiative forcing at the tropopause. Results reported by Wang et al. (1991), however, showed that, because of the differences in the vertical distribution of the radiative forcing, the calculated temperature responses do not scale as the net forcing at the tropopause. The temperature response to each greenhouse gas must be calculated on a case by case basis along with any indirect effect so that their climate impact can be judged in the context of global change.

The need for explicit model simulations are even more important for studies of the climatic

effects of ozone and stratospheric aerosols, which have been identified as having important implications for climate change (IPCC, 1994). Ramaswamy et al. (1992) first pointed out that the observed depletion of ozone over the past decade may have off-set the radiative forcing due to the accumulation of CFCs during the same period. Subsequent works (Molnar et al., 1994; Wang et al., 1993) have corroborated these results. Ozone redistribution is also expected from natural solar variability (see results from recent calculations in Brasseur, 1993 and Fleming et al., 1994); human activities such as use of industrial halocarbons (see e.g. WMO, 1995), the operation of subsonic aircraft (Johnson et al., 1992; WMO, 1994), and stratospheric aircraft (HSRP, 1993). The ozone distribution and its response to perturbations are often strong functions of latitude. A multi-dimensional photochemical model is needed to predict ozone changes. The standard tools for predicting ozone changes include 2-D chemistry-transport models with prescribed circulation and temperature, and interactive 2-D models that calculate the stratospheric temperature, heat transport, and transport circulation in a self-consistent manner. The latter models can take into account O₃ increases in the stratosphere caused by stratospheric cooling due to CO₂ increases (e. g., Haigh and Pyle, 1982; Schneider et al., 1993). The behavior of O₃ in the lower stratosphere is very sensitive to the tropospheric heating rates adopted in the models (Hou et al., 1991). An accurate accounting of the ozone redistribution must include the changes in tropospheric heating as the energy budget is perturbed. It seems necessary to study both greenhouse warming and (related) ozone-depletion effects as they occur in the real world, i. e. simultaneously in a fully interactive model.

Previous studies of the climatic effects of stratospheric aerosols have concentrated on how changes in aerosol loading may affect the radiative balance (see, e.g., recent work by Lacis et al. 1992). Changes in sulfate loading can also affect the rate of heterogeneous reactions and the ozone distribution in the lower stratosphere (see, e.g., Rodriguez et al., 1994; Pitari and Rizi, 1993; Granier and Brasseur, 1992; Tie et al., 1995). The effect from ozone redistribution should be evaluated so that the direct radiative effect of the aerosol can be isolated. The changes in radiative forcing arising from changes in ozone and stratospheric aerosols are expected to be a strong function of latitude. Specific climate and climate-chemistry feedback mechanisms must be examined in the context of a multi-dimensional model to analyze the response of the surface temperature to these forcings.

The global warming potential (GWP) should be representative of the total greenhouse impact of a trace gas. Previous calculations (Fisher et al., 1990) obtained the direct effect by calculating the time integrated change in radiative forcing resulting from the addition of the trace gas from the time it is introduced to the atmosphere to the time it is removed from the atmosphere. In the calculation, it is assumed that the atmospheric concentrations of all other trace gases are held fixed. In reality, the presence of a trace gas can also perturb the concentrations of other trace

gases. Such changes may also provide additional (negative or positive) radiative forcing. Such effects should be included as part of the indirect effect. For example, the radiative forcing that arises from the O₃ redistribution attributed to emissions of a CFC should be included as an indirect greenhouse effect of the CFC (see, e.g., Daniel et al., 1995). The IPCC (1994) report includes other examples, such as the indirect effect arising from water vapor and CO₂ production from CH₄. Thus such effects should be examined using models that can simulate the climate-chemistry interaction.

We contend that an interactive climate-chemistry model is the appropriate tool to examine these issues. Figure 1 shows the current configuration of our interactive climate-chemistry (ICC) model. It consists of two submodels, the AER Seasonal Radiation Dynamics (ASRD) model and the AER Chemistry Dynamics (ACD) model, joined through a coupling scheme. Predictions of climate changes in response to emissions of greenhouse gases and aerosols require the following steps: identifying the forcing mechanisms, predicting of the changes in concentrations of all greenhouse gases and aerosols from the forcing mechanism, calculating the changes in radiative forcing from changes in the concentrations, and predicting changes in temperature from changes in the forcing. Each of these steps is discussed below.

Figure 1: Schematic diagram showing the components of the AER interactive climate-chemistry (ICC) model. It consists of two submodels, the AER Seasonal Radiation Dynamics (ASRD) model and the AER Chemistry Dynamics (ACD) model, joined through a coupling scheme.

At this point, the IPCC reports have identified changes in concentrations of greenhouse gases (CO_2 , N_2O , CH_4 , CFCs, and O_3), aerosols, and variability in solar cycle as important factors affecting climate. We will explain below how the AER ICC model keeps track of the changes caused by each of these forcing functions.

a) Changes in Concentrations of Greenhouse Gase, Aerosols and Ozone

(i) Long-lived Greenhouse Gases: For trace gases with long atmospheric lifetimes (e.g. N_2O and the CFCs), changes in concentrations due to a one-time pulse emission can, in most cases, be related to the amount emitted through their atmospheric lifetimes in terms of simple analytic expressions. Methane is removed by reaction with OH in the troposphere. At the same time, the OH concentration in the troposphere depends on the CH_4 concentration. This coupling leads to a non-linear response of the CH_4 concentration to the emission flux (Sze, 1977), and an added amount of CH_4 actually is purged from the atmosphere more slowly than suggested by its atmospheric lifetime (Prather, 1994). The case of CO_2 is more complicated because of the coupling to the deep ocean and the biosphere. The ACD module in our ICC model calculates the evolution of all trace gases that are removed by in situ atmospheric reactions, wet deposition and surface deposition. It does not calculate CO_2 changes.

(ii) Changes in Aerosol Loading: Aerosol particles in the lower stratosphere consist of sulfuric acid/water droplets. The background concentration is maintained mainly by photo-oxidation of OCS, with DMS, H_2S , CS_2 , and anthropogenic SO_2 contributing a minor portion. Volcanic eruptions can introduce large amounts of SO_2 into the lower stratosphere, providing an additional source for the sulfate. The sulfate particles have a residence time of one to two years in the lower stratosphere.

We have developed a sulfate aerosol model that can calculate the size distribution and composition of stratospheric sulfate particles from specified sources of sulfur gases released at the ground or injected in the stratosphere by volcanic eruptions or high-flying aircraft. In future calculations, the results from this model can be used to calculate the changes in heterogeneous reaction rates in the ACD module and changes in optical properties of aerosol in the climate module.

(iii) Ozone: Emission of many of the greenhouse gases also leads to changes in O_3 (see, e.g., Ko et al., 1993). It has been generally accepted that emissions of CFCs are responsible for the decrease in the column abundance of O_3 during the past decade. Other studies suggest that increases in CH_4 could lead to higher concentrations of O_3 in the troposphere. Previous studies using 2-D models (Brasseur and Simon, 1981; Garcia et al., 1984; Callis et al., 1985) provided quantitative estimates for the sensitivity of ozone to long-term variations in solar flux at ultraviolet (uv) wavelengths. Results from four 2-D models containing gas-phase chemistry only that were

reported in WMO (1990) indicate that the global ozone content is 2% larger at solar maximum than at solar minimum. Results from models with heterogeneous chemistry are available from several recent studies. Unfortunately, it is difficult to compare the results because each work used different assumptions about the variation of the solar flux. Model calculated differences in column ozone between solar maximum and solar minimum range from 0.5% to 3.5%. These values are to be compared with the value of 1.2% derived from the statistical analysis of the Dobson data on ozone and the F_{10.7} solar flux through 1984 (Reinsel et al., 1987).

Rind et al. (1990) showed that both the vertical and latitudinal structure of the temperature change following a doubling of CO₂ results in increases in the propagation of planetary waves into the stratosphere, and increased potential energy in the lower stratosphere. These processes generated an increase in the eddy energy in the middle atmosphere with an attendant increase in wave forcing and residual circulation, tending to warm high latitudes. This change in the transport circulation should affect the ozone distribution.

The ACD module in the ICC model has the capability to calculate the ozone changes from the mechanisms discussed above.

b) Changes in Radiative Forcing

The changes in radiative forcing for the greenhouse gases are calculated based on the model-computed distribution of the gases as a function of latitude and altitude. As pointed out by Wang et al. (1991), the forcings from long-lived greenhouse gases such as CFC-11 and CFC-12 are different from that of CO₂. The bulk of the ozone depletion in the past decade is observed to have occurred below the (approximately) 25 km level. According to radiative forcing calculations (Wang and Sze, 1980; Lacis et al., 1990; Ramaswamy et al., 1992), decreases in ozone in the lower stratosphere at mid-latitudes lead to a decrease in radiative forcing at the tropopause, while decreases in ozone above 25 km lead to an increase in forcing.

The effect of stratospheric aerosol on radiative forcing was summarized in a recent paper by Lacis et al. (1992). The optical depth, single scattering albedo, and asymmetry factor of the aerosol layers can be calculated using Mie theory. These parameters can then be used in a radiative transfer code. The climatic effect of aerosols can be manifested in several ways. First, an increase in atmospheric aerosol concentration can decrease the amount of sunlight reaching the earth's surface by either reflecting more solar radiation back into space or by absorbing more solar radiation. Both processes will result in a cooler surface. However, cooling resulting from an increase in solar energy absorbed by the atmosphere will be compensated for somewhat, because the atmosphere will become warmer and subsequently radiate more thermal radiation downward toward the surface. In addition, larger aerosols, with sizes comparable to infrared wavelengths, can also effectively absorb terrestrial radiation and hence will act much as a greenhouse gas does.

Aerosol size distribution and amount, vertical distribution in the atmosphere, and chemical composition are all potentially important to determining the net radiative forcing of aerosols.

c) Climatic feedback for steady-state response to change in forcing

The complex relationship between surface temperature change and a given radiative forcing manifests itself in a myriad of feedback. As the surface temperature responds to the imposed forcing, many physical quantities will also change. These include moisture content, cloud cover, cloud optical properties, and behavior of sea-ice. Changes in these quantities will modify the initial forcing and the temperature response. The large range in model-predicted temperature responses is largely due to the different methods adopted in treating these feedbacks. Based on the results from a number of GCM models, Cess et al. (1990) estimated that the range of predicted global mean surface temperature increases due to CO₂-doubling is between 1.5°C and 4.5°C.

Apart from the issues of water vapor and cloud feedbacks, several groups have noted the possible link between climate change and changes in the thermohaline circulation of the deep ocean (see for example Watts and Morantine, 1990; Nakamura et al., 1994; and Broecker, 1990). Watts and Morantine developed a two-dimensional model of the ocean with eddy mixing in the vertical and horizontal directions, and deep-water formation at high latitudes, driving a mean circulation. They coupled their model to a simple one-layer energy balance model of the atmosphere similar to that described by North et al. (1981). Using this model they showed that a rapid reduction in the rate of deep water formation resulted in a surface cooling of up to 6°K at high latitudes. They argued that their results were consistent with the Younger Dryas reversal in which the deglaciation process at the end of the last ice age reversed and ice sheets reappeared again for a short time. Broecker and Denton (1990) have suggested that a rapid decrease in deep water formation could be linked to fresh water intrusion into the North Atlantic, decreasing the density of the surface water and disrupting ocean circulation.

d) Transient Response of the surface temperature to changes in radiative forcing

Broecker and Denton (1990) note that these rapid changes in ocean circulation may have occurred on time scales of less than 20 years. This suggests that the ocean circulation may also play a crucial role in the climate signal of the next 10 to 100 years. Both the large sensitivity of the surface temperature to deep ocean circulation changes and the short time scales over which these changes can take place suggest that this important mechanism should be investigated in greater detail to better understand the potential response of the climate system in the next decade to century. We plan to modify the model to include the effect of a mean circulation in the Ocean.

A single characteristic response time of the climate system is sometimes used in assessments of the relative influence of trace gas perturbations to global climate change. Choosing an effective heat capacity C of the system is equivalent to identifying the effective thickness of the ocean for interactions with the atmosphere over the time scales of interest 1-100 years. Harvey and Schneider (1985) used a 20-layer advection diffusion (UD) model of the coupled ocean-atmosphere system to show that some aspects of the transient response of the climate system to a step increase in radiative forcing cannot be captured by this single equivalent layer assumption. Based on qualitative considerations of the transient response of the HS UD model, Harvey (1986) suggests that it is more appropriate to characterize the transient response of the climate system with three time scales: the times to reach 25, 50, and 75 percent of the equilibrium response. Wigley and Raper (1987) developed a similar UD model to investigate the transient response of global mean surface temperature and sea level. Their model, or versions of it, has been used extensively in global change assessment studies (see for example, IPCC [1990]) section 6.6, and Wigley [1995]).

1.2 Highlights of scientific results from this project

A large portion the effort in this project was devoted to developing appropriate models to address the issues discussed in the previous section. The development of the models will be discussed in section 2 of this document. Here, we highlight the scientific results that were obtained by applications of these models.

(a) Ozone as a greenhouse gas

The three-zone AER Seasonal Radiation Dynamics (ASRD) model was used to calculate the changes in radiative forcing and steady state surface temperature for two different ozone depletion profiles that are consistent with the observed changes in total column ozone over the last decade. Depending on the assumed vertical distribution of the ozone depletion, the calculated ozone forcing offsets 17% and 35% of the decadal greenhouse forcing. When the changes in surface temperature are calculated, the offsets correspond to 28% and 69% of the surface warming. Our results showed that the calculated radiative forcing from changes in ozone depends on the vertical profiles of the ozone depletion and that the surface cooling due to lower stratospheric ozone loss is larger than the cancellation in forcing. The results were published in *JGR* and a copy is included in Appendix C.

These experiments have been repeated using the nineteen-zone ASRD model. Our results show that ozone depletion at high latitudes results in a cooling of the surface-troposphere system that is greater (by a factor of 2.8) than that estimated from the change in radiative forcing resulting

from ozone depletion and the model's $2\times\text{CO}_2$ climate sensitivity. Our analyses suggest that changes in meridional heat transport from low to high latitudes combined with the decrease in the infrared opacity of the lower stratosphere are very important in determining the steady state response to high latitude ozone losses. The 39% compensation in greenhouse warming resulting from lower stratospheric ozone losses is also larger than the 28% compensation simulated previously by the lower resolution model. The higher resolution model is able to resolve the high latitude features of the assumed ozone perturbation that are important in determining the overall climate sensitivity to these perturbations. The results were submitted for publication to the Journal of Climate. A copy of the manuscript is included as Appendix D.

(b) Climate-Chemistry Coupling

We used the Interactive Climate Chemistry (ICC) model to simulate the present day atmosphere and the doubled- CO_2 atmosphere. There is an increase in stratospheric ozone concentration in the double- CO_2 atmosphere because the stratosphere is cooler. The increase in ozone from double CO_2 acts as a negative feedback to moderate the stratospheric cooling compensating for 30% of the expected cooling around 45 km in the sunlit hemisphere. An increase in column content of 3.5% (1.5% at the tropics and 4.5% at high latitudes) is caused by the CO_2 -doubling. Diagnostic calculations show that the effect on ozone is mainly from the change in temperature. Additional details of the results are given in section 2 and in Appendix E.

The calculated ozone response is not significantly different from that calculated in Schneider et al. (1993). One of the primary reasons for this is that both models presently have a relatively weak exchange of energy, momentum, and chemical composition between the troposphere and stratosphere. Thus, calculating tropospheric temperatures interactively has a small influence in our present model framework. We suggest that incorporating a three-wave model structure into the ICCM will provide a more realistic simulation of stratospheric wave activity and possibly an improved representation of stratosphere-troposphere exchange.

(c) Feedback mechanisms affecting the steady state temperature response

We used the three-zone ASRD model to study the effects of low-latitude oceanic cloud/climate feedback in relation to trace gas increases. Two feedbacks have been postulated to limit the sea surface temperature (SST) response. One is the "cirrus thermostat" (proposed by Ramanathan and Collins, 1991) which postulates that the formation of highly reflective cirrus clouds will reduce the amount of solar radiation, limiting SST to temperatures no greater than about 305K. The other, proposed by Fu *et al.* (1992), favors evaporative (surface latent) heat flux changes acting as a "thermostat". Our calculation showed that the increase of cirrus fractional cover and cloud top height with increased deep convection could induce either a strong positive or

a negative feedback for SSTs, depending on cirrus microphysical properties. In particular, large cirrus particles would lead to a strong positive feedback, whilst small cirrus particles induce a negative feedback. The results of the analysis were published in the proceedings of an 1993 AMS conference. A copy is included as Appendix A in this document.

Available evidence suggests that the earth has never fully glaciated even when the sun was about 20-30% fainter than today. This is referred to as the young sun paradox (see e.g., Endal and Schatten, 1982; Crowley and North, 1991). The three-zone ASRD model was used to see whether feedback defined by the parameterizations for meridional heat transport and sea-ice can prevent complete glaciation. It is found that dynamical heat-flux feedback alone may have protected the Archean earth against a runaway glaciation. The result, published in *J. of Glaciology*, is included as Appendix B.

(d) Effect of ocean delay in the transient temperature response

We developed a 5-box global upwelling diffusion (UD) model (one near-surface atmosphere interacting with 4 ocean layers) of the coupled atmosphere-ocean system to estimate the transient response of the surface temperature to a given atmospheric forcing. Our analyses showed that the transient response of the surface temperature to an applied radiative forcing from changes in concentration of a greenhouse gas can be represented by five normal modes. The dominate modes representing 18%, 37%, 20% and 21% of the equilibrium response have time constants of 1.7 years, 7.2 years, 40 years, and 990 years respectively. The results will be submitted for journal publication and is included in this report as Appendix F.

The formalism was used to calculate the transient response of the surface temperature to changes in forcing due to expected changes in concentrations of the CFCs, HCFCs, HFCs, N₂O, CH₄ and CO₂ over the next century. The relative contribution to the incremental warming between 1950 and 2050 is shown to be dominated by increases in CO₂. Additional results can be found in appendix G.

1.3 List of Publications from this Project

Molnar, G. I. (1993) *Greenhouse sensitivity to tropical water vapor distribution and cirrus property changes*. Proceedings from the Fourth Symposium on Global Changes Studies, 17-22 January 1993, Anaheim, CA, Published by the American Meteorological Society, Boston, MA.

- Molnar, G.I., M.K.W. Ko, S. Zhou, and N.D. Sze (1994) Climatic consequences of observed ozone loss in the 1980s: Relevance to the greenhouse problem. *J. Geophys. Res.*, **99**, 25,755-25,760, 1994.
- Molnar, G.I. and W.J. Gutowski, Jr. (1995) The "faint young Sun paradox": further exploration of the role of dynamical heat-flux feed backs in maintaining global climate stability. *Journal of Glaciology*, **41**, No. 137, 87-90.
- MacKay, R.M., M.K.W. Ko, R.-L. Shia, and Y. Yang (1995) Ozone's contribution to the climate signal of the 1980s, Abstract No. A51B-11 for the American Geophysical Union 1995 Spring Meeting.
- Mackay, R.M., M.K.W. Ko, S. Zhou, G. Molnar, R-L Shia, and Y. Yang (1996) An estimation of the climatic effects of ozone during the 1980s: The role of the ozone hole. Submitted to *Journal of Climate*.
- MacKay, R.M., M.K.W. Ko, R-L Shia, S. Zhou, and Y. Yang (1996) The response of a coupled climate-chemistry model to greenhouse forcing. Manuscript in preparation.
- MacKay R.M. and M.K.W. Ko (1996) Normal Modes and the transient response of the climate system. Manuscript in preparation.
- Ko, M.K.W., R.M. MacKay, R-L Shia and N-D Sze (1996) An Estimate of global warming from chlorofluorocarbons and their alternatives over the next century. manuscript in preparation.

2. PROGRESS ON CURRENT PROJECT

The latest three-year funding cycle for the grant was scheduled to run from December 1992 to November 1995. Because of changes in personnel in the project at AER, we requested and received a no-cost extension for the project through August 1996.

This section summarizes the tasks accomplished during this period. The work during the past 44 months consisted of the following parts:

(1) Analysis of the results from the AER three-zone seasonal climate model

The three-zone AER seasonal climate (ASRD) model represents the globe using three latitude zones: the tropics between 30°S and 30°N and the extratropics in each hemisphere. The development of this model was completed prior to December 1992. We continued to analyze the model results during the early stage of the present project.

(2) Development of the Interactive Climate Chemistry model

A major goal of the project is to develop a climate-chemistry model to study climate-chemistry interactions. Recognizing that a finer meridional resolution is needed to resolve the spatial distributions of the chemical species, the first step in the development was to increase the meridional resolution of the ASRD model to 19 latitude belts with resolution of 10°. The nineteen-zone ASRD model was then coupled with the AER interactive chemistry dynamics (ACD) module to obtain the Interactive Climate-chemistry (ICC) model.

(3) Analysis of the transient response of global mean temperature to radiative forcing

The transient response of the surface temperature to applied radiative forcing depends on how the ocean circulation redistributes the incident energy. Many ocean-GCM's are being used to study this problem. We used a multi-box ocean-atmosphere model to derive the normal mode response of the atmosphere to applied forcing. The information should help us incorporate a more realistic ocean transport module in our ICC.

2.1 Progress in the first 36 months of the project

(a) Analyses of results from the three-zone AER Seasonal Radiation Dynamics (ASRD)

model

Role of Tropical Convective Adjustment: We performed simulations using the "standard" 6.5 K/km lapse rate, a moist-adiabatic lapse rate (with boundary-layer parameterization added), as well as employing a cumulus-convection parameterization (Lindzen *et al.*, 1982; Satoh and Hayashi, 1992) in the tropical zone. The last approach is the most interactive, because the vertical

temperature structure is not governed by a prescribed lapse rate. We found that although the surface temperature changes are similar for the moist adiabatic lapse rate simulation and the tropical cumulus parameterization, the cumulus convection scheme provides the largest mid-tropospheric warming, whilst moist adiabatic adjustment leads to the largest warming in the upper troposphere.

Cloud Optical Property Feedback: We implemented the Platt and Harshwardhan (1988) cirrus optical thickness parameterization into the code. Results (when only the tropical oceanic cirrus was parameterized this way) imply that, at least for the current climate, our choice of cirrus visible optical depth of 2 (at 0.55 mm) is appropriate.

Tropical Cloudiness/Sea-Surface Temperature(SST) Feedback: We used the ASRD model to study the effects of low-latitude oceanic cloud/climate feedback in relation to trace gas increases. It has been noted that the greenhouse trapping (difference between longwave radiance emitted to space, F and surface black-body radiance, E) increases sharply with rising SST. Defining the "Greenhouse Parameter" G , as $\Delta(E - F)/\Delta SST$, an apparent "super greenhouse effect" may be inferred (see, e.g. Ramanathan and Collins, 1991) indicating that over most of the tropical ocean ($SST > 297-300K$), G becomes very large (about $17-25 \text{ Wm}^{-2}\text{K}^{-1}$) compared to regions with lower SST values. Two feedbacks have been postulated to limit the SST response. One is the "cirrus thermostat" (proposed by Ramanathan and Collins, 1991) which postulates that the formation of highly reflective cirrus clouds will reduce the amount of solar radiation, limiting SST to temperatures no greater than about 305K. The other, proposed by Fu *et al.* (1992), favors evaporative (surface latent) heat flux changes acting as a "thermostat". Our calculation showed that the increase of cirrus fractional cover and cloud top height with increased deep convection could induce either a strong positive or a negative feedback for SSTs, depending on cirrus microphysical properties. In particular, large cirrus particles would lead to a strong positive feedback, whilst small cirrus particles induce a negative feedback. The results also indicate that evaporation changes alone, or even in combination with predicted wind speed changes, cannot bring the model-computed G close to its observed range. The results of the analysis were published in the proceedings of a 1993 AMS conference. A copy is included as Appendix A in this proposal.

The role of dynamical heat fluxes in maintaining global climate stability: Available evidence suggests that the earth has never fully glaciated even when the sun was about 20-30% fainter than today. This is referred to as the young sun paradox (see e.g., Endal and Schatten, 1982; Crowley and North, 1991). The three-zone ASRD model was used to see whether feedback defined by the parameterizations for meridional heat transport and sea-ice can prevent complete glaciation. It is found that dynamical heat-flux feedback alone may have protected the Archean earth against a runaway glaciation. The result, published in J. of Glaciology, is included as Appendix B.

(b) Studies on chemistry-climate interaction using the three-zone ASRD model

Radiative forcing due to ozone depletion in the past decade: The three-zone ASRD model was used to calculate the changes in radiative forcing and surface temperature for two different ozone depletion profiles that are consistent with the observed changes in total column ozone over the last decade. Depending on the assumed vertical distribution of the ozone depletion, the calculated ozone forcing offsets 17% or 35% of the decadal greenhouse forcing. When the changes in surface temperature are calculated, the offsets correspond to 28% or 69% of the surface warming. Thus, our results show that the calculated radiative forcing from changes in ozone depends on model details and the vertical profiles of the ozone depletion. The model-calculated surface temperature changes are steady-state responses. The "ocean-delay" effect would lessen the magnitude of the actual response. Nevertheless, when evaluating the ozone depletion response *relative* to the non-ozone gases cases only, the effects of ocean thermal inertia are much less pronounced (cf. IPCC, 1990). Our calculations show that the surface cooling due to lower stratospheric ozone losses is larger than the cancellation in forcing. The results were published in *JGR* and a copy is included in Appendix C.

Estimates of indirect GWP: We performed several calculations to estimate the indirect global warming potential (GWP) of a halogen compound and examined the results to determine whether they agree with those obtained with the scaling arguments using ozone depletion potential (ODP) values. In the calculations, we first used the AER ACD model to calculate the steady state concentration of a halogen species and the accompanying ozone depletion due to constant emission of the species. The 3-zone ASRD model was then used to calculate the changes in surface temperatures for two cases. In the first case, only the change due to an increase in the halogen concentration was imposed. The second case included the changes due to increases in the halogen and the accompanying decrease in ozone. The difference in the temperature responses was assigned as the indirect effect due to the ozone depletion. We performed the calculations for CFC-11 and HCFC-141b using the parameters in Table 2.1. For HCFC-141b, the model calculated cooling is smaller than the one estimated from its ODP.

Table 2.1: Estimate for indirect GWP

species	flux kiloton/year	cooling from ozone depletion per unit mass emitted (°K/kiloton)	calculated ratio of cooling per unit mass emitted	ratio of cooling per unit mass emitted using ODP
CFC-11	486	1.07×10^{-4}	1	1
HCFC-141b	2455	6.52×10^{-6}	0.063	0.090

(c) *Building the nineteen-zone ASRD model*

The nineteen-zone ASRD model includes 19 latitudinal zones from pole to pole. The first task we performed was to obtain two-dimensional distributions of prescribed model variables such as relative humidity, cloud cover (high, middle, and low), land albedo, and ozone concentrations (for uncoupled runs) for each month of the year.

The next task was to parameterize heat transport in the nineteen-zone ASRD model.

Heat transport in the stratosphere: In the first version of the nineteen-zone ASRD model, radiative equilibrium is assumed for the stratosphere. This results in winter temperatures in the lower polar stratosphere that are approximately 20°K too low. We have, therefore, included meridional heat transport in the stratosphere based on a temperature gradient driven turbulent diffusion.

Vertical heat transport in the troposphere: In the first version of the model, the vertical heat transport is implicitly included by using a baroclinic adjustment scheme in mid-high latitudes and a moist adiabatic adjustment scheme in the tropics (Stone and Carlson, 1979). In the current version, the lapse rate adjustment scheme has been modified slightly to improve the model's simulation of temperatures in the upper tropical troposphere. The moist adiabatic lapse rate (MAL) γ_m at the surface is used to calculate the rate of change of potential temperature (θ) with respect to pressure, p ,

$$\frac{\partial \theta}{\partial p} = \frac{0.288\theta}{p} \left(\frac{\gamma_m}{\gamma_d} - 1 \right) \quad (1).$$

with γ_d being the dry adiabatic lapse rate. The critical lapse rate, γ_c , at other levels is then calculated by assuming that $\partial\theta/\partial p$ is constant throughout the troposphere and inverting (1) with γ_c replacing γ_m . This moist critical lapse (MCL) rate is then compared with the critical lapse rate predicted by baroclinic instability theory and the actual critical lapse rate used is the smaller of the two.

Heat exchange between the land sector and the ocean sector: Because of the large difference in thermal inertia between ocean and land surfaces, separate land and ocean sectors are included within each latitudinal zone. The land and ocean heat exchange is computed according to the diffusion law with an empirical diffusion coefficient.

Parameterization of meridional heat transport (MHT): The meridional heat transport by the atmosphere is parameterized differently in mid-high latitudes and in the tropics. In mid-high latitudes, atmospheric heat transport is dominated by large-scale eddies. There have been a variety of parameterization schemes in the literature that are based on baroclinic theory. Here we adopt the scheme developed by Stone (1974) and modified by Peng et al. (1982). It is based in part on the Eady (1949) model and in part on the two-layer model (Holton, 1979). This scheme is somewhat different from that used previously in the 3-zone climate model (Wang et al., 1990), which was

developed by Branscome (1983) and based on the Charney (1947) model and quasi-geostrophic scaling analysis. We have evaluated and compared these physical schemes and another empirical scheme which is based on relations between total heat transport and temperature structure (North, 1975; Stone and Miller, 1980). In the tropics there are two important differences from mid-high latitudes. One is that heat transported by the atmosphere is smaller than that by the oceans. The other is that the eddy heat transport is much less than the Hadley cell transport. Unfortunately, a theoretical basis for parameterizing the Hadley cell heat transport is weak, and the empirical correlation between heat transport and temperature gradient is poor. It seems to us that the only practical scheme for an energy balance model, therefore, is that developed by Peng et al. (1987). This parameterization scheme is based on the theoretical results of Held and Hou (1980) and the empirical study of Rennick (1977). According to Peng et al., their simulation of tropical heat transport compares well to observations. Therefore, we use Peng et al.'s scheme in our model as a framework and make some modifications on the vertical distribution of heat transport. Peng et al. calculated the total heat transport first, then distributed it to each layer according to an empirical relation which is a third-order polynomial of surface temperature multiplied by the surface adiabatic lapse rate. To be consistent with parameterization schemes for vertical heat transport in our model, we do not use that empirical relation. Instead, we use moist adiabatic adjustment, which is a good approximation for the tropical atmosphere (Stone and Carlson, 1979).

Ocean heat transport: In the first version of the model, we used prescribed sea surface temperature (SST) from observations and use the model's surface energy imbalance to infer the sum of ocean heat transport and storage. The model was then modified to include a mixed 50 m deep layer ocean. The meridional heating of the mixed layer ocean is calculated assuming that the turbulent diffusion of ocean heat energy is directly proportional the meridional gradient of mixed layer temperature. Using the notation of Watts and Morantine(1990),

$$\frac{\partial T_o}{\partial t} = \frac{K_H}{(1 - f_i H_i / D) f_o R^2} \frac{1}{\cos \phi} \frac{\partial}{\partial \phi} \left(\cos \phi \frac{\partial T_o}{\partial \phi} \right) + \frac{(S + IR - LH - SH)(1 - f_i) + F_B}{(1 - f_i H_i / D) \rho C_w D} \quad (2)$$

where T_o is the mixed layer temperature, K_H the horizontal thermal diffusivity, R the radius of the earth, ϕ the latitude, $S+IR$ the net flux of solar plus terrestrial adiation into the ocean, LH and SH the latent and sensible heat fluxes from the ocean surface, ρ the density of water (taken as 1000 kg/m³), C_w the specific heat capacity of water (4186 J/kg/K), D the mixed layer depth (50 m), H_i the sea ice thickness, and F_B the rate of energy exchange (W/m²) between the ocean and sea ice. The term f_o , which is implicitly assumed to be unity in the study of Watts and Morantine, is the zonal fraction of ocean in a latitude belt, and f_i is the fraction of ocean that is covered by sea ice. We use constant values of $K_H=1.6 \times 10^{12}$ m²/yr and 2.1×10^{12} m²/yr for the northern and southern

hemispheres respectively. These are comparable to the value of $1.2 \times 10^{12} \text{ m}^2/\text{yr}$ used by Watts and Morantine (1990) and was adjusted to give a good simulation of observed ice cover and ocean temperature in each hemisphere. Figure 2 shows the model calculated meridional heat transport (in the ocean, in the atmosphere, and the sum) calculated for the present day atmosphere compared with observations reported by Peixoto and Oort (1992). The model simulates the magnitude and distribution of the total meridional heat flux reasonably well; i.e., peak fluxes at mid latitudes agree to within about 15% of observations. The difference between the simulated and observed meridional heat fluxes is largest for the ocean component. The simulated peak poleward oceanic heat transport is shifted poleward by 10 to 15 degrees relative to that of the observations in both hemispheres. Trenberth and Solomon (1994) quoted an estimate of the uncertainty in total oceanic heat transport of $\pm 0.45 \text{ PW}$ for 45°N . Our simulated oceanic heat transport does fall within this range of observation for this location.

Figure 2. Model simulated meridional heat transport in Oceans (short dash) Atmosphere (long dash) and Total (Atmosphere plus Oceans, solid) compared with observed values given by Peixoto and Oort (1992) for total transport (solid triangles), and by Trenberth and Solomon (1994) for oceanic transport (open triangles). Units are 10^{15} Watts (PW).

Thermodynamic sea-ice model: The thermodynamic sea-ice model used is that described by Molnar et al. (1994), except that here ocean fluxes are calculated as described in equation (1) instead of using the "q-flux" flux adjustment method. The ocean heat transport interacts directly with the sea-ice model and strongly influences the steady state cycle of sea-ice volume. This method of coupling ocean heat transport and sea-ice dynamics has a stronger physical basis than the prescribed heat flux method previously used in the three-zone ASRD.

(d) Coupling of the ACD module to the climate module

The two modules are complementary in that the climate module has better tropospheric representation (e.g., higher vertical resolution, more complete physical mechanisms, and land-ocean separation), while the ACD module has a better stratospheric representation (e.g., detailed transport of trace gases and chemistry). Because the physics and numerics of the two modules are very different and independent, we use a "modular" method to couple them. This method allows the two modules to run in parallel and to exchange information continuously during time integrations. In particular, we input the ACD module results of stratospheric temperature and gas concentrations to the climate module during the time integration, because in the climate module stratospheric temperature is in radiative equilibrium and gas concentrations are fixed. On the other

hand, we input the climate module results of tropospheric temperature to the ACD module, because in the ACD module tropospheric processes are only roughly represented. This information-exchanging procedure is a mutual data assimilation process. The fresh input data interact with other quantities and affect the whole model, making the two modules perform as an organic integral.

The exchange of temperature between the two modules requires some special consideration. In the stratosphere we simply replace the temperature field in the climate module with the temperature calculated from the ACD module. In the troposphere of the ACD module, however, we cannot directly replace the temperature field because the model's stratosphere is very sensitive to a change in lower boundary condition. Instead, we use an alternative approach. Because Newtonian cooling is used in the troposphere of the ACD module, with climatological temperature (Cunnold et al., 1975) used as the "target" temperature, we use the temperature calculated from the climate module to replace the climatological "target" temperature. This is equivalent to forcing the tropospheric temperature gradually toward that calculated by the climate module.

2.2 Progress in the previous 8 months of the project

(a) Simulations using the nineteen-zone ICC model

Molnar et al. (1994) used the three-zone ASRD to evaluate the steady state climatic consequences of sustained forcing similar in magnitude to the observed increases in greenhouse gases and ozone loss in the 1980s. These experiments have been repeated using the nineteen-zone ASRD model. The simulated increase in surface air temperature resulting from non-ozone greenhouse gases is 0.272 K. When changes in lower stratospheric ozone are included, the greenhouse warming is 0.165 K which is approximately 39% lower than when ozone is fixed at the 1979 concentrations. Ozone perturbations at high latitudes result in a cooling of the surface-troposphere system that is greater (by a factor of 2.8) than that estimated from the change in radiative forcing resulting from ozone depletion and the model's $2\times\text{CO}_2$ climate sensitivity. Our results suggest that changes in meridional heat transport from low to high latitudes combined with the decrease in the infrared opacity of the lower stratosphere are very important in determining the steady state response to high latitude ozone losses. The 39% compensation in greenhouse warming resulting from lower stratospheric ozone losses is also larger than the 28% compensation simulated previously by the lower resolution model. The higher resolution model is able to resolve the high latitude features of the assumed ozone perturbation which are important in determining the overall climate sensitivity to these perturbations. These results were submitted for publication to the Journal of Climate. A copy of the manuscript is included as Appendix D.

We used the ICC model to simulate the present day atmosphere and the doubled-CO₂ atmosphere. The model calculated temperature and the column abundance of ozone for the present day atmosphere are shown in figures 3 and 4 respectively, along with observations for comparison. The calculated change in temperature for CO₂-doubling is shown in figure 5a. In the calculation, there is an increase in stratospheric ozone concentration because the stratosphere is cooler. The increase in ozone from double CO₂ acts as a negative feedback to moderate the stratospheric cooling. Figure 5b shows the calculated warming in December plotted in a scale to better resolve the stratosphere. The warming that can be attributed to the accompanying ozone increase is shown in figure 5c. The results show that the compensation is quite significant around 45 km in the sunlit hemisphere. The increase in column ozone caused by the CO₂-doubling is shown in figure 6a. The ICCM model predicts an increase in stratospheric mean mass residual circulation of up to 10 %. Figures 6b and 6c show results from diagnostic calculations to quantify the effects on ozone from the changes in temperature and circulation. The results indicate that the change in circulation has a relatively small effect. Additional details of the results are given in Appendix E.

The calculated global increase of 3.5% is slightly greater than the 3.1% calculated in the chemistry/dynamics model (Schneider et al., 1993). The calculated change in circulation is larger than that calculated in Schneider et al. (1993). However, the differences between the two simulations are relatively small. One of the primary reasons for this is that both models presently have a relatively weak exchange of energy, momentum, and chemical composition between the troposphere and stratosphere. Thus, calculating tropospheric temperatures interactively has a small influence in our present model framework. We suggest that incorporating a three-wave model structure into the ICC model will provide a more realistic simulation of stratospheric wave activity and possibly an improved representation of stratosphere-troposphere exchange.

Fig. 3a. Zonal mean temperature profile simulated by the 19-zone ASDR for annual, JJA, and DJF means. Contour intervals are 10 K and height is approximate.

Fig. 3b. Zonal mean observed temperature profile for annual, JJA, and DJF means. Contour intervals are 10 K. Observations from NCAR Data Support Section, Scientific Computing Division.

Figure 4a: Column ozone amount for the present day atmosphere simulated by the ICC Model.

Figure 4b: Climatological mean values of the column ozone amount as derived from the TOMS measurement.

Figure 5b. The 2CO₂-1CO₂ temperature change for December. Values are plotted using the chemistry/dynamics grid to better highlight the stratosphere. Units are K

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Figure 5c. Temperature changes from the ozone redistribution that accompanies the CO₂-doubling for December). Units are K

Figure 5a. The calculated two-dimensional 2CO₂-1CO₂ temperature changes for DJF, JJA, and Annual means. Units are K and temperature changes less than or equal to zero are shown by dashed lines. The climate module grid is used for this plot to better highlight the troposphere.

(a)

(b)

(c)

Figure 6: (a) The percent change in column ozone as a function of latitude and month that results from the CO₂-doubling for the fully interactive run, (b) Percent change in column ozone as a function of latitude and month for the temperature change alone, and (c) Percent change in column ozone as a function of latitude and month for the change in mean mass stream function (dynamics) alone.

(b) Normal mode response of the surface temperature to applied radiative forcing

We developed a 5-box global upwelling diffusion (UD) model (one near-surface atmosphere interacting with 4 ocean layers) of the coupled atmosphere-ocean system to estimate the transient response of surface temperature to a given atmospheric forcing. The results from the model are similar to the 20-box UD model of Harvey and Schneider (1985). The UD model is solved to yield the time constants of the five natural decay modes of the 5-box system and the contribution of each mode to the equilibrium response to forcing arising from changes in atmospheric concentrations of greenhouse gases. Our analyses showed that the transient response of the surface temperature to an applied radiative forcing from changes in concentration of a greenhouse gas can be represented by five normal modes. The dominant modes representing 18%, 37%, 20% and 21% of the equilibrium response have time constants of 1.7 years, 7.2 years, 40 years, and 990 years respectively. The results will be submitted for journal publication and is included in this report as Appendix F.

This formalism is used to calculate the transient response of the surface temperature to changes in forcing due to expected changes in concentrations of the CFCs, HCFCs, HFCs, N_2O , CH_4 , and CO_2 over the next century. The calculated surface warming relative to the pre-industrial era for scenario IS92a from IPCC is shown in figure 7. The relative contribution to the incremental warming between 1950 and 2050 is shown in figure 8. In both cases, increases in CO_2 contribute the most to the realized warming. Additional results can be found in appendix G.

IS92a- Warming relative to 1750

Figure 7: The Calculated surface warming relative to the pre-industrial era due to increases in greenhouse gases as specified in the IPCC IS92a scenario.

Figure 8: The contribution of each greenhouse gas to the incremental warming between 1950 and 2050 according to the IPCC IS92a scenario.

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