INCREASED ATMOSPHERIC CARBON DIOXIDE AND CLIMATE FEEDBACK MECHANISMS

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INTRODUCTION

As a consequence of fossil fuel burning, the atmospheric concentration of carbon dioxide has increased from 314 ppm in 1958, when detailed measurements of this quantity began, to a present value of 335 ppm; and it is estimated that during the next century, the CO$_2$ concentration will double relative to its assumed preindustrial value of 290 ppm. Since CO$_2$ is an infrared-active gas, increases in its atmospheric concentration would lead to a larger infrared opacity for the atmosphere which, by normal logic, would result in a warmer Earth. A number of modeling endeavors suggest a $2^\circ$ to $4^\circ$C increase in global mean surface temperature with doubling of the CO$_2$ concentration.

But such estimates of CO$_2$-induced warming are highly uncertain because of a lack of knowledge of climate feedback mechanisms. Interactive influences upon the solar and infrared opacities of the Earth-atmosphere system can either amplify or damp a climate-forcing mechanism such as increasing CO$_2$. This paper discusses a number of such climate feedback mechanisms.

CLIMATE SENSITIVITY

A convenient measure of the sensitivity of the global climate, for the purpose of feedback comparisons, is the sensitivity parameter $\beta = S_0 \frac{dT_S}{dS}$ as originally introduced by Schneider and Mass (1975), where $T_S$ is the global mean surface temperature and $S$ the solar constant, with $S_0$ denoting the current solar constant. Thus $\beta$ is a measure of the sensitivity of global climate to a change in solar constant. From a global energy balance,

$$\frac{S}{4}(1 - \alpha_p) = F$$

where $\alpha_p$ and $F$ denote the global albedo and outgoing infrared flux, respectively. It thus follows from equation (1) that

$$\beta = S_0 \frac{dT_S}{dS} = \frac{F}{dF/dT_S} + (S_0/4)(d\alpha_p/dT_S)$$

\[ \text{(2)} \]
The absence of climate-induced changes in either infrared opacity or albedo of the Earth-atmosphere system, which comprise the feedback coupling mechanisms of present concern, results in \( \frac{d\alpha}{dT_g} = 0 \), while \( \frac{dF}{dT_s} \) is evaluated as follows. Let \( F = \varepsilon_0 T_g^4 \), where \( \varepsilon_0 \) is the Stefan-Boltzmann constant; then for \( F = 233 \text{ W/m}^2 \) and \( T_s = 288 \text{ K} \), \( \varepsilon = 0.6 \) is the emissivity of the Earth-atmosphere system. Thus \( \frac{dF}{dT_s} = 3.3 \), and from equation (2),

\[
\beta = 70^\circ C
\]  

so that a 1-percent change in \( S \) would produce a 0.7°C change in \( T_g \). The above constitutes a reference value for the sensitivity parameter, in that it contains the basic temperature-radiation negative feedback but none of the possible feedback associated with variable optical properties of the Earth-atmosphere system.

Change in water vapor amount is a positive atmospheric feedback mechanism since an increase in surface temperature increases the water vapor content of the atmosphere, which increases the atmospheric infrared opacity with a subsequent further increase in surface temperature. Most studies are in agreement that this feedback roughly doubles climate sensitivity. For example, Cess (1976) has empirically suggested that \( \frac{dF}{dT_g} = 1.6 \) for which \( \beta = 145^\circ C \), essentially twice that of equation (3).

An additional positive mechanism is ice-albedo feedback. A warmer planet results in less snow and ice cover and thus in a lower albedo and in turn a yet warmer planet. Annual models (Lian and Cess, 1977; Manabe and Wetherald, 1975) suggest that this mechanism increases \( \beta \) from 145°C to 185°C, although a recent seasonal model (Wetherald and Manabe, 1981) indicates that an annually averaged seasonal model is more stable than the analogous annual model; this subject is discussed further in a later section.

In addition to water vapor and ice-albedo feedbacks, there are numerous other possible feedback mechanisms, perhaps the most controversial being cloudiness-radiation feedback. In the following section, some aspects of this mechanism are discussed.

CLOUDINESS-RADIATION FEEDBACK

Cloudiness-radiation feedback contains two uncertain aspects. The first is the question of whether or not cloud amounts, heights, optical properties, and structure will significantly change in response to CO₂-induced warming. If such changes are not significant, then obviously there will be no cloudiness-radiation feedback. But if cloud amounts and heights are influenced by climatic change, then both the solar and the infrared components of the radiation budget will be altered; it is the relative role of these radiation changes that constitutes the second uncertain aspect of the problem.

For example, if cloud amounts decrease, then since clouds are generally brighter than clear-sky regions, the Earth-atmosphere system albedo would be reduced, resulting in increased solar heating of the system. But decreased
cloudiness would also reduce the infrared opacity of the atmosphere, resulting in increased infrared cooling of the Earth-atmosphere system. Thus the separate solar and infrared modifications act in opposite directions. A corresponding change in effective cloud height would further modify the outgoing infrared radiation: a reduction in effective cloud height, for example, would enhance infrared cooling since the lower (and hence warmer) clouds would radiate more energy to space.

Employing a general circulation model which predicts both cloud amount and cloud height, Manabe and Wetherald (1980) have suggested that equatorward of 50° latitude, doubling and quadrupling atmospheric CO₂ would reduce net cloud amount and effective cloud height because of CO₂-induced warming; both effects increase the outgoing infrared radiation. But this increase is nearly compensated in the model by the corresponding increase in absorbed solar radiation due to reduced cloud amount. Poleward of 50°, they found an increase in net cloud amount without any substantial change in effective cloud height. But the absence of the latter, which contributed to the near solar-infrared compensation at lower latitudes, is in effect offset by reduced insolation at higher latitudes, so that again the model predicts near compensation between absorbed solar and outgoing infrared radiation.

Manabe and Wetherald (1980) emphasize, "In view of the uncertainty in the values of the optical cloud parameters and the crudeness of the cloud prediction scheme incorporated into the model, it is premature to conclude that the change of cloud cover has little effect on the sensitivity of climate." There have, in fact, been suggestions (Petukhov et al., 1976; Hunt, 1981; Wang et al., 1981; Charlock, 1981) that changes in cloud optical properties associated with climatic change might be important in modeling cloudiness-radiation feedback.

Alternate approaches to estimating the relative solar-infrared components of cloudiness-radiation feedback involve empirical studies using Earth radiation budget data. In one such approach, Cess (1976) has suggested solar-infrared compensation, whereas Ohring and Clapp (1980) and Hartmann and Short (1980) have suggested that the solar component dominates the infrared component by roughly a factor of 2. Cess employed the satellite data compilation of Ellis and Vonder Haar (1976), while the other two studies utilized radiation budget data derived from scanning radiometer measurements by the National Oceanic and Atmospheric Administration (NOAA). Recently Cess et al. (1982) have reviewed these studies and illustrate that such empirical conclusions depend substantially upon the satellite data set which is employed. They further suggest that the conclusions of solar dominance (Ohring and Clapp, 1980; Hartmann and Short, 1980) might be attributable to the NOAA albedo values at 0900 local time together with the NOAA data being derived from narrow spectral measurements.

Clearly the empirical approaches comprise an important means of studying the cloudiness-radiation feedback problem. The approach by Ohring and Clapp (1980) is particularly attractive. They have employed interannual variability in regional monthly mean radiation data, from which they estimate the relative solar-infrared cloudiness feedback components by attributing this variability to interannual variability in cloudiness. It would seem most worthwhile to reexamine their conclusions by employing radiation budget data which do not suffer the possible deficiencies noted above.
CLIMATE CHANGE PREDICTIONS

An obvious test of a climate model is its ability to reproduce seasonal climatic change. Recently Ramanathan et al. (1979) have formulated a seasonal energy balance climate model; the seasonal change in surface air temperature predicted by this model is compared with observations in figure 1. The extremely good agreement between predicted and observed surface air temperatures is due in part to tuning of the model, since the latitudinal heat capacity of the Earth-atmosphere system was tuned to the observed surface air temperature. Nevertheless, tuning a single parameter results in both phase and amplitude being correct.

Reproducing the observed seasonal cycle does not, however, ensure that the model can correctly predict a perturbed climate. For example, the energy balance model does not account for changes in the heat capacity of the Earth-atmosphere system due to changes in the thickness and extent of sea ice. But Wetherald and Manabe (1981) have indicated that such changes can significantly influence the high-latitude seasonal response of surface temperature to enhanced atmospheric CO$_2$. Furthermore, Wetherald and Manabe find that their seasonal model predicts significantly less sensitivity to increased atmospheric CO$_2$ than does the corresponding annual model, and this also appears to be related to changes in sea ice thickness and extent.

To illustrate the dependence of climate sensitivity upon high-latitude Earth-atmosphere heat capacity changes associated with sea ice changes, we have changed arbitrarily high-latitude heat capacity within the seasonal energy balance model described by Ramanathan et al. (1979). This change is summarized in table I for a doubling of atmospheric CO$_2$. Here $R$ is the latitudinal heat capacity of the Earth-atmosphere system, with the values for present CO$_2$ being those obtained by tuning to the present climate. For doubled CO$_2$, we have arbitrarily increased $R$ at the high latitudes to crudely mimic the fact that a warmer planet would result in a reduction in both thickness and extent of sea ice, and thus allow more efficient interaction between atmosphere and ocean with a corresponding increase in the heat capacity of the Earth-atmosphere system at these latitudes.

Figure 2 illustrates the seasonal increase in surface air temperature in several high-latitude regions for doubled CO$_2$ concentrations, both with and without the change in $R$ listed in table I. Clearly our arbitrary change in $R$ is not realistic, since at 85° N, it reduced annual mean surface air temperature. But the point of figure 2 is obvious: the perturbed seasonal cycle at high latitudes is highly sensitive to climate-induced changes in $R$. Moreover, the qualitative features shown in figure 2 are consistent with the model results of Wetherald and Manabe (1981), in that maximum high-latitude warming occurs during the winter. This is also consistent with an interesting empirical study by Vinnikov and Groysman (1979). Employing observed climatic change over roughly the past century, they have evaluated $\Delta T_S/\Delta T_S$ as a function of season and latitude, where $\Delta T_S$ is the change in seasonal surface temperature at a given latitude and $\Delta T_S$ is the corresponding change in global mean surface temperature. Their results are summarized in figure 3, from which it is again shown that maximum sensitivity occurs during the winter.
Aside from zonal seasonal sensitivity, recall that Wetherald and Manabe (1981) additionally found that global warming due to increased atmospheric CO$_2$ was reduced when they employed a seasonal rather than an annual climate model. The present energy balance model suggests that this reduction is associated with the influence of the change in $R$ upon the seasonal cycle. Employing the energy balance model in both annual and seasonal modes, we obtain the following increases in global mean surface temperature for a doubling of atmospheric CO$_2$ concentration:

$$\Delta T_S = 3.3^\circ C \text{ for annual model}$$
$$\Delta T_S = 3.3^\circ C \text{ for seasonal model with fixed } R$$
$$\Delta T_S = 2.7^\circ C \text{ for seasonal model with variable } R$$

The point here, of course, is that the seasonal model produces reduced global warming only if the heat capacity is allowed to vary.

In addition to warming as a consequence of CO$_2$ additions to the atmosphere, fossil fuel burning can alter chemical composition in other ways through interactive atmospheric chemistry; this is discussed in the following section.

**INTERACTIVE ATMOSPHERIC CHEMISTRY**

Recent atmospheric chemical models indicate that as a consequence of increasing emissions of CO, NO$_x$, and CH$_4$ due to fossil fuel burning, tropospheric ozone and methane concentrations might significantly increase in the near future (Logan et al., 1978; Hameed et al., 1979). This increase in tropospheric O$_3$ and CH$_4$ is a consequence of interactive chemical processes involving species derived from CH$_4$, H$_2$O, NO$_x$, and O$_2$. Although only about 10 percent of atmospheric ozone is located within the troposphere, this ozone contributes roughly half of the total ozone greenhouse effect because of pressure broadening of the 9.6-$\mu$m band (Ramanathan and Dickinson, 1979). Thus an increase in tropospheric ozone, in conjunction with a corresponding increase in tropospheric methane, could possibly produce a significant climatic effect, augmenting the global warming due to increasing atmospheric CO$_2$, which is also a consequence of fossil fuel burning.

An additional interactive effect is that a warmer climate would, by itself, influence atmospheric chemical composition, since increased atmospheric water vapor, resulting from a warmer climate, would produce increased OH, an important constituent in chemical reactions which govern the amounts of tropospheric ozone and methane. Thus, not only can changes in atmospheric composition alter the climate, but also climatic change can alter atmospheric composition through interactive chemistry. This then constitutes a climate feedback mechanism.

To crudely appraise whether or not increased emissions of CO, NO$_x$, and CH$_4$, resulting from increased fossil fuel consumption, could significantly augment the related CO$_2$ warming, we have constructed a coupled climate-chemical model for the purpose of investigating the sensitivity of the global climate to changes in CO, NO$_x$, and CH$_4$ emissions. This model is described as follows.
The climate-chemical model is a combination of the energy balance climate model described by Ramanathan et al. (1979) and an extended version of the tropospheric chemical model of Hameed et al. (1979). Both models employ latitude as the sole dimension. The climate model incorporates separately calculated changes in surface-troposphere radiative heating due to changes in atmospheric CO\(_2\) and tropospheric O\(_3\) and CH\(_4\). The chemical model employs vertically averaged conditions, so that it predicts changes in the total tropospheric column abundance of O\(_3\) and CH\(_4\) as a function of latitude.

Figure 4 illustrates, in schematic form, the most important chemical processes resulting from changes in CO, CH\(_4\), and NO\(_x\) emissions. An increase in CO emission will, for example, lead to a conversion of OH to HO\(_2\). Since OH is the only known tropospheric sink for CH\(_4\), this reduction in OH thus leads to an increase in tropospheric CH\(_4\). In addition, the increased HO\(_2\) converts NO to NO\(_2\), with photodissociation of the latter producing odd oxygen which then combines with O\(_2\) to increase the level of tropospheric O\(_3\). Thus, increasing the emission of CO leads to increases in both CH\(_4\) and O\(_3\).

An increase in CH\(_4\) emission does, of course, directly increase the concentration of CH\(_4\). Moreover, the oxidation of CH\(_4\) produces CO which, by the process just discussed, increases the concentration of O\(_3\).

Increasing the emission of NO\(_x\) also leads to an increase in O\(_3\) as a consequence of the NO\(_2\) to O to O\(_3\) process just discussed. In addition, photodissociation of O\(_3\) produces O\(^{1}\text{D}\) which reacts with H\(_2\)O to form OH; then the increased OH subsequently leads to a reduction in CH\(_4\). So, although enhanced NO\(_x\) emission leads to increased tropospheric O\(_3\), increased O\(_3\) reduces the amount of tropospheric CH\(_4\). Of course, the dependence of OH upon O\(_3\) also modifies the increase in CH\(_4\) due to increasing either CH\(_4\) or CO emissions.

Changes in tropospheric O\(_3\) and CH\(_4\), through altering the infrared opacity of the troposphere, lead to climatic change, and this in turn influences the atmospheric composition through a secondary feedback loop as illustrated in figure 4. For example, an increase in both tropospheric CH\(_4\) and O\(_3\) would lead to warming of the Earth-tropospheric system, which in turn would increase tropospheric H\(_2\)O and, through the reaction with O\(^{1}\text{D}\), lead to enhanced OH. This in turn would decrease the CH\(_4\) concentration, as well as the concentration of CO, and subsequently of O\(_3\). Thus the climate-chemical interaction results in a negative feedback mechanism, partially, but not totally, reducing the increased CH\(_4\) and O\(_3\) which produced climate change in the first place. Obviously there are numerous interactive processes at work concerning the influence of chemical composition change upon climate.

To model the coupled climatic change, the CO\(_2\)-climate model of Ramanathan et al. (1979) is employed. In that model, climatic change was induced by reducing the outgoing latitudinal infrared flux by the amount \(\Delta F(\text{CO}_2)\) corresponding to a specified increase in atmospheric CO\(_2\). The quantity \(\Delta F(\text{CO}_2)\) was determined from a separate radiation calculation. We use the same procedure in this study, with \(\Delta F(\text{CO}_2)\) replaced by

\[
\Delta F = \Delta F(\text{CO}_2) + \Delta F(\text{O}_3) + \Delta F(\text{CH}_4) \tag{4}
\]
The calculations of $\Delta F(O_3)$ and $\Delta F(CH_4)$ account for the increase in infrared opacity of the troposphere due to the fundamental vibration-rotation bands of $O_3$ (9.6 µm) and $CH_4$ (7.8 µm). Ultraviolet (UV) absorption by $O_3$ within the troposphere has been neglected, since most of the UV is absorbed within the stratosphere. As in the comparable $CO_2$ calculation, overlap by water vapor absorption is included as well as the influence of clouds. Illustrative results for $\Delta F(O_3)$ and $\Delta F(CH_4)$, corresponding to a doubling of the present atmospheric concentrations of both species, are shown in figure 5.

It should be emphasized that the present climate-chemical model does not include an interactive stratosphere. Since the stratosphere acts as the primary source of tropospheric $O_3$, we would not anticipate changes in tropospheric $O_3$ to significantly influence the stratosphere. But just the reverse is the case for $CH_4$: tropospheric methane is transported into the stratosphere, so that changes in tropospheric $CH_4$ should induce stratospheric change.

Although carbon-cycle models exist for the purpose of estimating future atmospheric $CO_2$ concentrations, it is a far more hazardous endeavor to attempt to predict future $CO$, $NO_x$, and $CH_4$ emissions. Table II lists estimated anthropogenic and natural emissions of these quantities for the entire globe in 1976. Except for major depressions and wars, the annual increase in fossil fuel consumption has been 4.3 percent over the past century (Rotty, 1978). If this growth rate continues, fossil fuel consumption will increase by a factor of 8 in 2025 relative to 1976. But this by no means implies that anthropogenic emissions of $CO$, $NO_x$, and $CH_4$ will increase by the same factor.

Roughly 60 percent of the present anthropogenic $CO$ emission stems from automobiles, and it is not likely that future petroleum consumption by automobiles will increase at the same rate as total fossil fuel consumption. On the other hand, conventional coal-fired power plants produce twice the amount of $CO$ per unit of energy as do oil-fired plants, because of the combined effects of coal's lower heating value and its less complete combustion. Thus, in this context, conversion from oil to coal would by itself lead to increased anthropogenic $CO$ emission. But technological improvements in the utilization of coal, such as coal gasification and the use of fluidized bed reactors in the burning of pulverized coal, could reduce future $CO$ emission per unit of energy. In principle, of course, it is technologically possible to significantly reduce emissions of $CO$, $NO_x$, and $CH_4$; however, the global extent to which this might be done will surely be influenced by economic factors.

Future changes in natural emissions of $CO$, $NO_x$, and $CH_4$ are just as difficult to predict. Vegetation is the primary source of natural $CO$, while a substantial production of natural $CH_4$ stems from swamps and rice fields. Deforestation and the draining of swamps might decrease such emissions, although it has been suggested that deforestation may be leveling off as the result of more efficient use of existing agricultural land (Rotty, 1979). There is, in fact, some evidence for reforestation on the North American Continent, with forests replacing abandoned agricultural land which has proven to be only marginally productive. On the other hand, future $CO_2$ warming of the biosphere, coupled possibly with increased global precipitation and $CO_2$ fertilization, might tend to significantly increase natural $CO$ and $CH_4$ emissions. In particular, the natural production of $CH_4$ is strongly temperature dependent.
Obviously the above discussion precludes any sort of quantitative estimate of future CO, NO\textsubscript{X}, and CH\textsubscript{4} emissions. For the present purpose of illustrating nothing more than the sensitivity of global climate to changes in these emissions, we choose arbitrarily the following scenario:

1. The CO\textsubscript{2} content of the atmosphere increases by a factor of 1.7, consistent with Keeling and Bacastow's (1977) carbon-cycle model for 2025 relative to 1976.

2. Natural emissions of CO, NO\textsubscript{X}, and CH\textsubscript{4} remain the same.

3. Anthropogenic emissions of these constituents increase by factors of 4 and 8.

4. Latitudinal emission distributions remain the same.

Results of the coupled climate-chemical model, for this hypothetical 1976 to 2025 scenario for the change in emissions, are listed in tables III and IV. In table III, no account has been taken of the previously discussed negative climate-chemical feedback resulting from the climate-induced change in atmospheric H\textsubscript{2}O. Without changing the anthropogenic emissions of CO, NO\textsubscript{X}, and CH\textsubscript{4}, global temperature is increased by 2.63°C because of the increase in CO\textsubscript{2} concentrations by a factor of 1.7. Significantly greater warming occurs when CO\textsubscript{2}, NO\textsubscript{X}, and CH\textsubscript{4} emissions are increased.

The results of table IV incorporate the negative climate-chemical feedback. For no change in anthropogenic CO, NO\textsubscript{X}, and CH\textsubscript{4} emissions, increasing OH resulting from global warming due to the CO\textsubscript{2} increase reduces tropospheric O\textsubscript{3} and CH\textsubscript{4} by factors of 0.9 and 0.85, respectively. The resulting climatic effect is rather minor, reducing the prior 2.63°C warming to 2.45°C. Although reduced somewhat from the corresponding values in table III, the increase in hemispheric warming due to increased emissions of CO, NO\textsubscript{X}, and CH\textsubscript{4} is still significant. Relative to the global temperature increase of 2.63°C, which corresponds to no interactive chemistry, global warming is increased by 0.7°C and 1.4°C for increases in anthropogenic emissions by factors of 4 and 8, respectively.

We again emphasize that we chose the assumed 1976 to 2025 scenario regarding the increase in CO, NO\textsubscript{X}, and CH\textsubscript{4} emissions solely for the purpose of illustrating how the climate responds to a change in such emissions. A realistic appraisal of future emissions of these constituents requires knowledge not only of future power-production technology on a global scale, but also of the manner in which the biosphere responds to future climatic change. It might well be that the latter is more significant than the former. The conclusion of this study is simply that the climate can respond to changes in the total emissions of CO, NO\textsubscript{X}, and CH\textsubscript{4}, as the result of the ensuing changes in tropospheric O\textsubscript{3} and CH\textsubscript{4}. 
REFERENCES


Rotty, Ralph M. 1979: Uncertainties Associated With Global Effects of Atmospheric Carbon Dioxide. ORAU/IEA-79-6(0), Oak Ridge Assoc. Univ.


Wang, Wei-Chyung; Rossow, William B.; Yao, Mao-Sung; and Wolfson, Marilyn 1981: Climate Sensitivity of a One-Dimensional Radiative-Convective Model With Cloud Feedback. J. Atmos. Sci., vol. 38, no. 6, pp. 1167-1178.

TABLE I.- VALUES OF HEAT CAPACITY AT HIGH LATITUDES FOR PRESENT AND DOUBLED ATMOSPHERIC CO₂ CONCENTRATIONS

<table>
<thead>
<tr>
<th>CO₂ concentration</th>
<th>Heat capacity R, J/m²°C, at latitudes of</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>45° N</td>
</tr>
<tr>
<td>Present</td>
<td>33</td>
</tr>
<tr>
<td>Doubled</td>
<td>33</td>
</tr>
</tbody>
</table>

aR values in model tuned to present climate.

TABLE II.- ESTIMATED 1976 EMISSIONS OF CO, NOₓ, AND CH₄

<table>
<thead>
<tr>
<th>Constituent</th>
<th>Anthropogenic emission, metric tons/year</th>
<th>Natural emission, metric tons/year</th>
</tr>
</thead>
<tbody>
<tr>
<td>CO</td>
<td>700 × 10^6</td>
<td>2730 × 10^6</td>
</tr>
<tr>
<td>NOₓ</td>
<td>20</td>
<td>16</td>
</tr>
<tr>
<td>CH₄</td>
<td>95</td>
<td>380</td>
</tr>
</tbody>
</table>
TABLE III.- HYPOTHETICAL 1976 TO 2025 CLIMATE CHANGE SCENARIO
WITHOUT CLIMATE-CHEMICAL FEEDBACK

<table>
<thead>
<tr>
<th>Anthropogenic emissions of CO, NO\textsubscript{X}, and CH\textsubscript{4} increased by factors of -</th>
<th>Resulting factor increases in -</th>
<th>$\Delta T_S$, °C</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$O_3$</td>
<td>$CH_4$</td>
</tr>
<tr>
<td>0</td>
<td>1.00</td>
<td>1.00</td>
</tr>
<tr>
<td>4</td>
<td>1.68</td>
<td>1.92</td>
</tr>
<tr>
<td>8</td>
<td>2.29</td>
<td>3.62</td>
</tr>
</tbody>
</table>

TABLE IV.- HYPOTHETICAL 1976 TO 2025 CLIMATE CHANGE SCENARIO
WITH CLIMATE-CHEMICAL FEEDBACK

<table>
<thead>
<tr>
<th>Anthropogenic emissions of CO, NO\textsubscript{X}, and CH\textsubscript{4} increased by factors of -</th>
<th>Resulting factor increases in -</th>
<th>$\Delta T_S$, °C</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$O_3$</td>
<td>$CH_4$</td>
</tr>
<tr>
<td>0</td>
<td>0.90</td>
<td>0.85</td>
</tr>
<tr>
<td>4</td>
<td>1.50</td>
<td>1.53</td>
</tr>
<tr>
<td>8</td>
<td>2.00</td>
<td>2.68</td>
</tr>
</tbody>
</table>
Figure 1.- Comparison between observed and calculated seasonal zonal surface temperatures.
Figure 2. - Calculated change in zonal seasonal surface temperature $\Delta T_s$ for doubling of CO$_2$ concentration, with and without change in heat capacity $R$. 
Figure 3. - Empirically determined zonal seasonal climate sensitivity \( \frac{\Delta T_s}{\Delta T_s} \) from Vinnikov and Groysman (1979).
Figure 4.- Schematic illustration of the major chemical reactions affecting tropospheric O$_3$ and CH$_4$.

Figure 5.- The surface-troposphere heating for doubling O$_3$ and CH$_4$ relative to their present concentrations.