On the Carbon Dioxide–Climate Confusion

STEPHEN H. SCHNEIDER

National Center for Atmospheric Research,[1] *Boulder, Colo. 80303*

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ABSTRACT

A number of estimates of global surface temperature sensitivity to a doubling of atmospheric carbon dioxide to 600 ppm are collected here and critically reviewed. The assumptions and formulations that lead to differences between certain models' estimates are explained in some detail. Based on current understanding of climate theory and modeling it is concluded that a state-of-the-art order-of-magnitude estimate for the global surface temperature increase from a doubling of atmospheric CO_2 content is between 1.5 and 3 K, with an amplification of the global average increase in polar zones. It is pointed out, however, that this estimate may prove to be high or low by several-fold as a result of climatic feedback mechanisms not properly accounted for in state-of-the-art models.

1. Introduction

Various estimates of the effect of changes in atmospheric carbon dioxide on global average surface temperature have been made. Many of these have subsequently been quoted by combinations of these same and other authors, generally in the context of warnings of potential global climatic effects of increasing carbon dioxide resulting from the burning of fossil fuels. As for these projected increases, Machta and Telegades (1974) have estimated a CO_2 increase of 20% by the year 2000; Bacastow and Keeling (1973) upped this to 25%, and they even project a doubling of CO_2 by 2040 if present trends continue; and Hoffert (1974) projects such a doubling of CO_2 to occur as early as 2025. That such increases in CO_2 might have an influence as large as natural climatic trends by 2000 has been postulated by many [e.g., recent references are Mitchell (1972), Kellogg and Schneider, (1974) and Broecker (1975)]. A useful surface temperature-CO_2 concentration sensitivity parameter, γ_2, can be defined which gives an estimate of globally-averaged surface temperature increase, ΔT_s, that is computed from various models for a doubling of the present value of atmospheric carbon dioxide, $[CO_2]$, from about 300 to 600 ppm; that is,

$$\gamma_2 \equiv \frac{\Delta T_s}{\Delta[CO_2]}[CO_2], \qquad (1)$$

where $\Delta[CO_2]=300$ ppm and $[CO_2]=300$ ppm for γ_2. Fig. 1 lists the various values of γ_2 found by a number of different authors from their models, although these models are often based on differing assumptions. The purpose of this paper is to explain briefly and *in one place* some major reasons for the differences between these estimates, in the hope of reducing some of the apparent confusion over these varying estimates that still seems to exist in much of the current literature despite the publication in several places of a few of the reasons for these discrepancies (e.g., Manabe and Wetherald, 1967, p. 250; Rasool and Schneider, 1971, footnote 12; Manabe, 1971; Sellers, 1974; Manabe and Wetherald, 1975).

2. Discussion of several models

Classical studies of potential CO_2 effects on climate were made by Chamberlin (1899) and Arrhenius (1903) and their ideas have given way to a plethora of follow-up studies. Plass (1961, among others) computed the surface temperature response of doubling CO_2 with a surface energy balance calculation. His earlier estimates were sharply contested by Kaplan (1961), who maintained that inclusion of cloudiness would reduce Plass' estimate considerably. Möller (1963) attempted to reconcile these conflicts, but heightened interest further by arguing that the atmosphere tends to conserve relative, rather than absolute, humidity; the latter assumption leads to his perplexing estimates for γ_2 seen on Fig. 1. However, all of these authors, though incorporating differing radiation models and atmospheric assumptions, shared one crucial assumption [as pointed out by Manabe and Wetherald (1967, p. 250)]: their surface temperature estimates were based on computations of changes in the *surface energy budget* primarily caused by the increased downward IR flux reaching the surface resulting from increased atmospheric IR opacity from increased CO_2; that is, they

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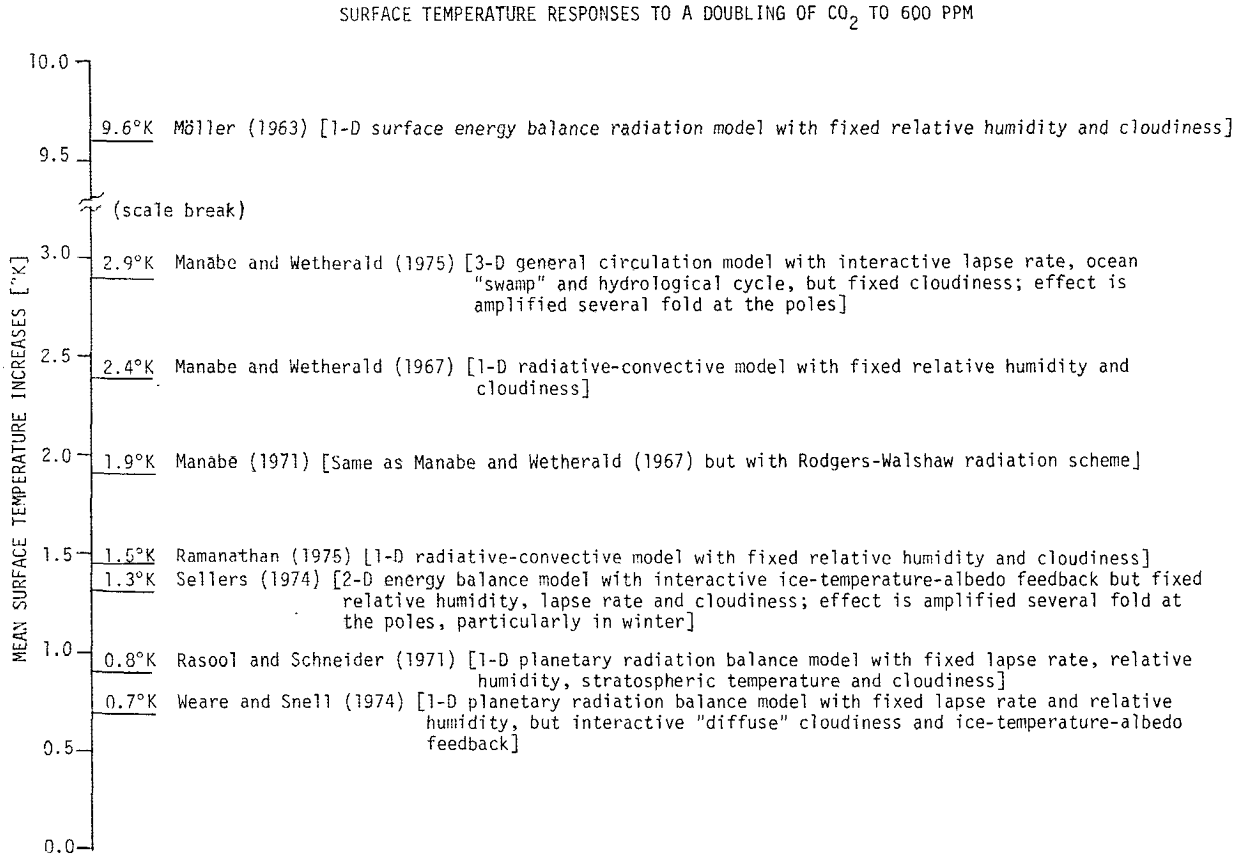

FIG. 1. Mean surface temperature sensitivity of various models to a doubling in the concentration of atmospheric carbon dioxide (γ_2). For the reasons discussed in the text a state-of-the-art order-of-magnitude estimate is suggested between 1.5 and 3 K, but that the combined effects of improperly modeled climatic feedback mechanisms could, roughly, enhance or reduce this estimate by as much as a factor of 4.

computed an equilibrium condition for the earth's surface rather than for the earth-atmosphere system as a whole. Manabe and Wetherald showed that none of these authors adequately included in their surface energy budgets the mixing effects of vertical heat transport by atmospheric motions. Manabe and Wetherald (1967) modeled this phenomenon by constructing a radiative-convective model; i.e., if the radiative calculation *per se* gave rise to a "supercritical" vertical temperature profile they then adjusted the lapse rate in the time evolution of their computations to a "critical" value ($= -6.5$ K km^{-1}). Since, on a *global average*, the radiative heating *per se* in the surface layers is sufficient to create a supercritical lapse rate (see Fig. 5 of Manabe and Wetherald, 1967), much of the increased downward IR flux at the surface from increased CO_2 would not remain as increased heating in the surface layer but rather would be carried upward by vertical motions and latent heat transport. Thus, the increased CO_2 would warm the entire globally averaged lower atmospheric temperature profile slightly, rather than dramatically changing the surface temperature as Möller's results suggest (and which can be seen in Fig. 1). However, the lower atmosphere is not vertically well-mixed everywhere, particularly in polar regions and mid-latitudes in winter. Thus, the global average radiative-convective calculations of Manabe and Wetherald (1967) underestimate the surface temperature warming effect of increased CO_2 in those regions of local vertical stability [as shown by Manabe and Wetherald (1975) with the GFDL general circulation model—as also described by Smagorinsky (1974)]. This is particularly important in polar regions, where the surface temperature increase from CO_2 was computed to be several times greater than the global average. Because the polar ice fields (sea ice in particular) are thought to be especially sensitive to changes in energy inputs at the surface, global CO_2 increases could have serious consequences in these regions (Kellogg, 1975; Budyko, 1972; SMIC, 1971).

Also, the effect of any change in surface energy budget components that causes a surface temperature response can, in turn, set off variations in other components which might "feedback" to modify the temperature change caused by the initial change. For example, Budyko (1969) and Sellers (1969, 1973, 1974) considered the self-amplifying effect of the feedback between snow and ice albedo and temperature, and incorporate parameterizations of this effect into their models. The results show that their formulations of this positive feedback could amplify the steady-state surface temperature response to a given perturbation to a component (e.g., the solar constant) of the energy balance by as much as a factor of 4. On the other hand, Paltridge (1974) and Weare and Snell (1974) have constructed models in which cloudiness-surface temperature interaction is represented; and their formulations (which are highly idealized, as are the ice feedback parameterization mentioned above) yield a negative feedback effect that reduces their models' surface temperature response to energy budget perturbations by factors of 2 or more. However, as Schneider and Dickinson (1974) point out in their survey of climate modeling, these individual feedback processes are not yet well enough understood (especially the effect of clouds) to ascertain whether the just-referenced attempts to model them are adequate as to the direction of the effect, let alone to determine whether the combination of these or other omitted processes would substantially modify the surface temperature sensitivity of a global-average radiative-convective model like that of Manabe and Wetherald (1967).

Therefore, since it is not known whether the synergism of the many feedback mechanisms operative in the climate system would amplify or dampen a perturbation to the surface temperature predicted by a simple globally-averaged radiative-convective model, perhaps the most reasonable *order of magnitude* estimate of the sensitivity of the surface temperature to increases in CO_2 that can be obtained from state-of-the-art knowledge still comes from a globally-averaged model that computes radiative processes in some detail and also accounts for vertical heat transport. Perhaps this is the reason the Manabe and Wetherald estimate ($\gamma_2=2.36$ K) remains the most widely quoted estimate of carbon dioxide-surface temperature sensitivity, γ_2. Yet other models that are globally averaged, accounting through lapse rate assumptions for convection and ignoring cloud or ice feedback mechanisms, give different results. For example, Rasool and Schneider (1971) computed $\gamma_2=0.8$ K, nearly a factor of 3 smaller than Manabe and Wetherald's estimate. A natural question is: why are these results different and who is "right?" The first part of this question is easier to answer than the latter, as we shall discuss next.

Manabe and Wetherald have a different (but also non-interactive) cloudiness prescription from Rasool and Schneider and a different (but very similar) water vapor profile [although both sets of workers agree with Möller (1963) that constant *relative* humidity is a preferable global average assumption to constant *absolute* humidity]. Visible albedo is held fixed in Rasool and Schneider but is somewhat interactive in Manabe and Wetherald since they include absorption of near-infrared solar energy by water vapor. Rasool and Schneider used separate (but fixed) tropospheric and stratospheric lapse rates, but Manabe and Wetherald can

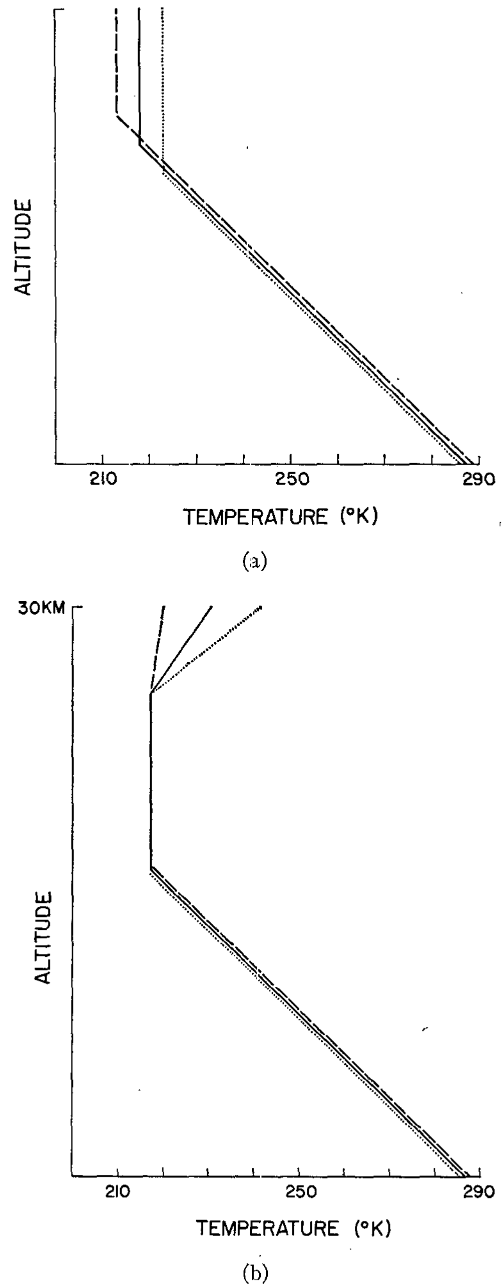

FIG. 2. Schematic diagram of the vertical temperature profiles used to study the ratio of change in surface temperature to change in stratospheric temperature needed to maintain planetary radiation balance at the top of the atmosphere (after Coakley and Schneider, 1974). The solid curve represents the profile of the control experiment while the dotted curve represents the profile for the warmer stratosphere and the dashed curve represents the profile for the cooler stratosphere. In (a), if the entire stratosphere changes temperature by about 4 K, the troposphere will vary by 1 K in the opposite sense. In (b), if the 30 km stratospheric temperature changes by about 20 K, the troposphere will vary by 1 K in the opposite sense.

have any lapse rate provided no part of the vertical temperature profile exceeds a critical lapse rate. The IR schemes are also different: Manabe and Wetherald use an admittedly crude wavelength-integrated "emissivity" type formulation whereas Rasool and Schneider perform a wavelength integration over the IR spectrum that has many spectral intervals. The latter formulation, however, was (also admittedly) based on earlier absorption data than used by Manabe and Wetherald and on an exponential transmission function T_r of the form

$$T_r \propto e^{-Ku},$$

where K represents an absorption coefficient and u the amount of absorber.

The first question then is which of these fundamental differences in the two models accounts for the fact that Manabe and Wetherald predict a γ_2 some 1.6 K larger than Rasool and Schneider. Wetherald and Manabe have also been concerned with these questions and kindly agreed to rerun their γ_2 calculations with the assumptions of Rasool and Schneider. The chief differences can be traced to three points:

1) The increased CO_2 increases IR cooling to space in the stratosphere (which decreases the stratospheric temperature) and increases stratospheric IR opacity. Since Rasool and Schneider maintain constant stratospheric temperature even after CO_2 is doubled, their model's stratosphere emits too much infrared radiation to space. This, combined with the constraint of planetary radiation balance at the top of the atmosphere, results in a compensation process by which their model's troposphere does not warm sufficiently. This effect was modeled by Manabe and Wetherald and not by Rasool and Schneider, and, as explained in the next section, accounts for roughly 0.5 K of the 1.6 K difference in γ_2 estimates.

2) The decrease in the earth's albedo for near IR solar radiative from increased water vapor and CO_2 concentrations leads to an effect that can account for several more tenths of a degree K. This effect too was included by Manabe and Wetherald and not by Rasool and Schneider.

3) A more elaborate treatment of infrared radiation transfer reduces Manabe and Wetherald's estimate for γ_2 by about 0.5 K. In fact, Manabe (1971) obtained this result when he used a Rodgers and Walshaw (1966) formulation for computation of IR fluxes in the previous Manabe and Wetherald model. Manabe (1971) obtained $\gamma_2=1.9$ K in this case, nearly a half degree less than γ_2 for Manabe and Wetherald.

3. Physical explanation of the two different model sensitivities

a. Planetary radiation balance model

The effect mentioned in point 1) can be partially understood from Fig. 2 (from Coakley and Schneider, 1974). The figures show the relation between an arbitrary stratospheric temperature change and the companion tropospheric temperature change that is required to maintain planetary radiation balance *at the top of the atmosphere* in a one-dimensional radiative energy balance model, but with CO_2 unchanged. Fig. 2a shows that planetary radiation balance requires $\sim$4 K uniform stratospheric temperature decrease to be accompanied by a $\sim$1 K surface temperature increase. [A similar relationship can be inferred from Fig. 12 of Manabe and Wetherald (1967), although this situation is for a change in stratospheric composition whereas our Fig. 2 is for the hypothetical case of fixed stratospheric composition.] If the lower stratospheric temperature is unchanged up to 70 mb, but increases linearly upward as in Fig. 2b, then the ratio of stratospheric cooling at 30 km to surface warming is about 20 to 1. Fig. 16 of Manabe and Wetherald (1967) suggests that a doubling of CO_2 decreases stratospheric temperature by about 5 K at 30 km, with the decrease in temperature growing with height. Wetherald and Manabe have used their original radiative convective model to find, as pointed out earlier, that neglect of stratospheric temperature decrease and downward IR flux increase from CO_2 doubling leads to an underestimate of γ_2 by about 0.5 K. This is in the range of what we would expect if the relationships from Fig. 2a and 2b are taken as extremes.

Thus we have seen that for fixed composition and a stratospheric temperature that is too warm (cold), the troposphere will be too cold (warm). Therefore the Rasool and Schneider model, which did not reduce stratospheric temperature with increased CO_2, gives too much IR flux to space from the stratosphere, hence too little IR upward flux is emitted from the troposphere—and it undergoes too small a temperature change.

These figures are for the case of unchanged composition and the constraint of planetary radiation balance at the top of the atmosphere. But of course, the composition of the stratosphere is changed when CO_2 is doubled, so in the interest of obtaining better physical insight to the role of stratospheric opacity and temperature change with CO_2 increase let us take a further look at the tropospheric response to the stratospheric temperature and compositional change from doubling of CO_2.

b. Radiative-convective equilibrium model

In addition to the constraint of maintaining planetary radiation balance at the top of the atmosphere, a radiative-convective equilibrium model also computes local temperatures from differencing upward *and* downward radiation fluxes.

When the IR opacity of the stratosphere is increased from a doubling of CO_2 the stratosphere will then emit more infrared radiation down toward the lower atmosphere (thereby warming the troposphere) and out to space (thereby cooling the stratosphere). However, the resultant cooling of the stratosphere reduces the magnitude of IR fluxes that would be emitted both up to space and down toward the tropopause. Thus, there are two competing effects on $F_T^{\downarrow}$, the downward IR flux at the tropopause, that occur when the CO_2 amount is increased *and* the stratospheric temperature is decreased: (i) an increase in $F_T^{\downarrow}$ from the increased IR opacity owing to the increased CO_2 and (ii) a decrease in $F_T^{\downarrow}$ from the decreased stratospheric temperature.

To obtain an estimate of the relative importance of these competing effects we use an improved version of the model in Coakley and Schneider (1974). The model is used to compute $F_T^{\downarrow}$ for three cases:

- A control case (CO_2=300 ppm) in which $F_T^{\downarrow}=0.0202$ cal cm^{-2} min^{-1}.
- A case in which CO_2 is doubled and the stratospheric temperature is allowed to change (i.e., cool) until radiative convective equilibrium is re-established in which $F_T^{\downarrow}=0.0217$ cal cm^{-2} min^{-1}, a value 0.0015 cal cm^{-2} min^{-1} larger than $F_T^{\downarrow}$ for the control case. This could be called the "stratospheric greenhouse effect," perhaps. Thus, despite the fact that the stratosphere was cooler, the CO_2 doubling case still produced a greater downward IR flux at the tropopause than the control case, indicating that the increased IR opacity effect dominates the decreased stratospheric temperature influence.
- To put these two opposite effects in relative perspective, a third case is computed in which the cooler stratospheric temperatures calculated from the CO_2 doubling case are used, but the control amount of CO_2 (i.e., 300 ppm) is retained. For this case $F_T^{\downarrow}=0.0195$ cal cm^{-2} min^{-1}, a value only 0.0007 cal cm^{-2} min^{-1} less than the control value of $F_T^{\downarrow}$ and slightly less than half of 0.0015 cal cm^{-2} min^{-1}, which is the difference between $F_T^{\downarrow}$ of the control case and $F_T^{\downarrow}$ of the CO_2 doubling case.

In the Manabe and Wetherald radiative-convective equilibrium model (in which both upward and downward IR fluxes are computed) the stratosphere cooled as a result of the increase in CO_2 opacity, but the cooling did not fully offset the increase in downward IR flux from increased opacity. Thus, an accurate calculation of the increase in IR opacity and decrease in temperature of the stratosphere is required to obtain an accurate surface temperature sensitivity value, γ_2.

c. *Reconcilation between these approaches*

These two approaches could be reconciled if the stratospheric temperatures in the model of Rasool and Schneider were determined to give radiative equilibrium in the stratosphere; then the surface temperature sensitivity to a doubling of CO_2 would have been similar to that of a radiative-convective model *even though* there is no downward IR flux calculation in the former approach. The two methods would give similar results under these conditions because of the imposition of the planetary radiation balance condition at the top of the atmosphere, which would force the troposphere to warm up appropriately when the stratosphere were cooled. However, the radiative-convective approach (in which both upward and downward IR fluxes are calculated and stratospheric and tropospheric temperature profiles mutually adjust) would be necessary to evaluate the perturbed stratospheric temperature—which is a requisite input in order to get an accurate calculation of γ_2 from the planetary radiation balance model.

The lack of a stratosphere (and thus absence of stratospheric cooling) in Sellers (1974) probably accounts for the relatively low value of γ_2 he calculated with his model, which otherwise would have had a much larger value of γ_2 owing to the inclusion of ice-albedo feedback. Furthermore, the relative underestimate of γ_2 is probably even larger for Sellers than for Rasool and Schneider, since the latter authors' model includes the effect of increased stratospheric opacity on decreasing the transmission of tropospheric IR flux to space. Thus, the absence of a stratosphere would further diminish the increase in upward tropospheric IR flux needed to maintain planetary radiation balance when CO_2 is doubled.

Ramanathan (1975),[2] also using a one-dimensional radiative-convective model [but with different infrared and solar radiation formulations than Manabe and Wetherald (1967) or Rasool and Schneider (1971)], obtained a value of $\gamma_2=0.76$ K for the modeling assumptions of Rasool and Schneider [a description of his model has been published in Ramanathan (1974)]. Ramanathan then included the effect of stratospheric cooling and opacity increase and his γ_2 increased by 0.44 K to 1.2 K (almost the same as Wetherald and Manabe's 0.5 K increase). When Ramanathan extended his model to include solar albedo change due to CO_2 and H_2O absorption increases (i.e., similar to assumptions in Manabe and Wetherald) he obtained $\gamma_2 \approx 1.45$ K, an increase of 0.25 K due to albedo changes. In a private communication Wetherald and Manabe (1975) report that inclusion of albedo changes account for 0.27 K of their value for γ_2. Yet, despite the close agreement between Ramanathan and Wetherald and Manabe over the magnitudes of the contribution of points 1) and 2) of Section 2 to their computed values of γ_2, the latter authors' lowest value of γ_2 is 1.9 K and Ramanathan's value is $\sim$1.5 K. Adding the contributions of points 1) and 2) of Section 2 to Rasool and Schneider's original result also yields $\gamma_2 \approx 1.5$ K.

In view of the need for accurate computation of the CO_2 induced stratospheric temperature change (which

[2] Private communication. To be presented at Second AMS Conference on Atmospheric Radiation.

also may depend upon dynamical processes not included in these column models) and the different vertical placement of the non-interactive clouds in each of these models, it is difficult to explain unambiguously the roughly 0.5 K differences in γ_2 computed by these models; differences that remain even after the models incorporated similar assumptions about stratospheric temperature and opacity and absorption by H_2O and CO_2 of solar IR.

4. Conclusion

A simple one-dimensional globally averaged model of the earth-atmosphere system that accounts for mutual adjustments (through vertical heat transports of radiative, sensible and latent energies) among the stratosphere, troposphere and surface, and assumes a constant tropospheric relative humidity profile is, perhaps, the most reasonable available quantitative estimate of the sensitivity of the mean global surface temperature to increases in CO_2. A critical examination of the differences between the one-dimensional column models of Manabe and Wetherald (1967) and Rasool and Schneider (1971) suggests a global average surface temperature increase of some 1.5 to 2.4 K from a doubling of CO_2. However, extending the global average models to include the regions of relatively stable atmospheric stratification (i.e., polar regions) and the effect of snow-temperature-albedo feedback shows that these areas can have a greatly enhanced *surface* temperature response to increased CO_2 (Manabe and Wetherald, 1975). [It must be pointed out, however, that Manabe and Wetherald (1975) use the same IR scheme as Manabe and Wetherald (1967), rather than the Rodgers-Walshaw scheme used by Manabe (1971).]

This amplified surface temperature increase in polar regions from a doubling of CO_2 could, if eventually proved realistic, be a potentially crucial result, since the polar zones are believed to be regions of particularly high sensitivity to changes in energy balance parameters (e.g., see Budyko, 1972; Kellogg, 1975; Schneider and Dickinson, 1974). Perhaps, the most important consequence of a significant warming of the polar regions is the implication such a warming might have for changes in sea ice extent, glacial volume, or the world sea level. Unfortunately, quantitative estimates of such potential effects are still extremely difficult to make.

Thus, in the absence of more complete knowledge of the combined effects of various omitted feedback processes, in particular clouds (which could either dampen or amplify our order-of-magnitude estimate), we must rely essentially on the column models to provide an estimate of γ_2. Furthermore, if one added the enhanced sensitivity to CO_2 increases predicted to occur in the stably stratified polar regions (where snow-temperature-albedo feedback is also present) to the results of globally-averaged column models, a state-of-the-art order-of-magnitude estimate of the global surface temperature sensitivity to a doubling of atmospheric CO_2 to 600 ppm can be given: $1.5\ \mathrm{K}<\gamma_2<3\ \mathrm{K}$ (recognizing that in polar regions the response could be several times greater than the global estimate, particularly in winter). Furthermore, the combined effects of unknown or improperly modeled feedback mechanisms could modify this estimate by severalfold; but, as we have said, it is still impossible to determine even whether such feedbacks might enhance or reduce this estimate for γ_2.

One additional point should be made pertaining to the surface temperature sensitivity to changes in CO_2. Our sensitivity parameter γ_2 is defined for a doubling of CO_2, but surface temperature response (in a model without ice, cloud or other such feedbacks) to changes in CO_2 is not linear, but nearly *logarithmic* with CO_2 concentrations [as can be seen on Fig. 1 of Rasool and Schneider (1971) or Fig. 19.8 of Machta and Telegadas (1974)]. Thus, increases in CO_2 less than a doubling would cause (neglecting other feedbacks) a proportionately larger temperature increase than a mere linear scaling of CO_2 concentrations from γ_2 (similarly, changes greater than a doubling need to be scaled down from a linear extrapolation).

The important and perplexing dilemma posed by the present inability of climate theory and modeling to offer much more than an order-of-magnitude estimate of the climatic effects of increased CO_2 is that the seriousness of potential climatic risks of continued use (or social risks of abandoned use) of fossil fuel to the year 2000 and beyond range from negligible to extreme (e.g., as discussed by Schneider and Dennett, 1975), depending upon whether one believes that the collective influence on the climate of all improperly accounted for climatic feedback mechanisms would enhance or dampen the prediction of radiative-convective column models. Since the consequences of a climate change at the higher end of the current estimate could be both enormous and possibly irreversible, perhaps society would be best to err conservatively in planning future fossil fuel consumption patterns—and in any case should consider what preparations need to be made to adjust to such a dramatic change (Schneider and Mesirow, 1976).

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