NOTE

Questions Concerning the Possible Influence of Anthropogenic CO_2 on Atmospheric Temperature

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ABSTRACT

Estimates of the atmospheric temperature changes due to a doubling of CO_2 concentrations have been with a static radiative flux model. They yield temperature changes >0.25 K. It appears that the much larger changes predicted by other models arise from additional water vapor evaporated into the atmosphere and not from the CO_2 itself.

1. Introduction

It is generally agreed that increasing atmospheric CO_2 will lead to higher atmospheric temperature, and global temperature changes corresponding to a doubling of CO_2 have been estimated at about 2–3 K (Manabe and Wetherald, 1975; Schneider, 1975; Augustsson and Ramanathan, 1977). The predicted temperature changes are largest at high latitudes, amounting to 7–10 K in the model by Manabe and Wetherald and it has been suggested that one of the first effects may be the melting of the west Antarctic ice sheet, with a concomitant 5 m rise in sea level, which would inundate many coastal cities and much of Florida and the Netherlands (Mercer, 1978). Such possible consequences cast the potential CO_2 temperature rise as one of the most important problems in applied meteorology today. It therefore seems justified to raise questions about the predictions from a number of different viewpoints and that is the purpose of the present note.

2. Why should the temperature increase?

The answer to this question is usually treated as obvious and is given as the "greenhouse effect." Leaving aside the possibility that this term is a misnomer (e.g., see Newell and Dopplick, 1970), the implication is that additional infrared radiation from the extra CO_2 in the air is returned to the earth's surface, thereby contributing to heating of the surface and thence to heating of the air by the usual surface processes of latent heat and sensible heat transfer. It is important to stress, however, that CO_2 is not the main constituent involved in infrared transfer. Water vapor plays the major role and ozone is also of importance. With the infrared region divided into 22 spectral intervals, the infrared and solar fluxes have been computed at levels from the surface up to 5 mb using a procedure originally developed by Rodgers (1967) and modified by Dopplick (1972). The procedure has previously been applied to the computation of heating rates for increased CO_2 concentrations (Newell and Dopplick, 1970; Newell *et al.*, 1972). Table 1 gives the results of computations using standard climatological data for January. Twenty of the spectral intervals are dominated by water vapor and the other two contain CO_2 ($\sim$15 μm) and O_3 ($\sim$9.6 μm), although overlap with water vapor is also included. Calculations were performed for CO_2 concentrations of 330 and 600 ppmv, taking care to include the changed CO_2 concentrations also in the near-infrared solar absorption (cf. Newell *et al.*, 1972). Both sets of computations were also repeated with cloud absent.

The infrared flux dominated by CO_2, as is well known, is only about 10% of that controlled by water vapor. The decrease in infrared flux from the surface to the atmosphere due to the increase in CO_2 ranges from 1.0–1.6 W m^{-2}. The increased CO_2 yields additional absorption of solar infrared radiation and therefore a decrease of solar energy available at the surface which ranges up to $\sim$0.3 W m^{-2}. The net change at the surface is an increase of 0.8–1.5 W m^{-2} with the smallest values

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TABLE 1. Energy flux at the surface for CO_2 concentrations of 330 and 600 ppmv with cloudy skies—January climatology. Units: W m^{-2}.

Latitude	Infrared flux in bands dominated by H_2O	CO_2 330 ppmv	CO_2 600 ppmv	O_3	Solar flux in near infrared CO_2 330 ppmv	CO_2 600 ppmv	Flux change 600–330 ppmv Infrared	Solar	Net change Cloudy	Clear
80°N	−34.79	−2.98	−1.74	−6.86	—	—	1.24	—	1.24	2.20
60°N	−31.83	−3.10	−1.99	−7.96	7.23	7.20	1.11	−0.03	1.08	2.24
40°N	−41.44	−4.10	−2.82	−12.62	32.41	32.30	1.28	−0.11	1.17	2.19
20°N	−52.76	−4.86	−3.35	−20.66	69.90	69.71	1.51	−0.19	1.32	1.70
0	−28.96	−3.66	−2.60	−14.22	84.61	84.43	1.06	−0.18	0.88	1.18
20°S	−30.11	−3.30	−2.25	−14.20	97.67	97.46	1.05	−0.21	0.84	1.15
40°S	−43.65	−4.15	−2.82	−15.90	103.12	102.85	1.33	−0.27	1.06	1.52
60°S	−37.59	−3.83	−2.69	−11.62	78.50	78.26	1.19	−0.24	0.95	1.85
80°S	−49.26	−5.54	−3.95	−11.81	32.00	31.89	1.59	−0.11	1.48	2.61

at low latitudes. When the calculations were repeated for clear skies, the net increase was 1.1–2.6 W m^{-2}.

The net increase in surface energy flux of $\sim$1 W m^{-2} at low latitudes may be compared with the net increase computed by Manabe and Wetherald of 3.5 W m^{-2}. Their estimate is thus substantially higher than ours; they do not break down the actual fluxes into contributions by CO_2 and H_2O and it is possible that the major part of their computed increase in infrared transfer actually originates from an increase in water vapor, which is predicted by their model to accompany an increase in temperature. Our radiative flux computations assume constant temperature.

Thus our answer to the current question is that there is indeed an increase in the energy received at the surface (reflected in a decreased net infrared loss from the surface) from the additional CO_2. It is not immediately evident that this same answer applied to the Manabe-Wetherald model as much of the extra energy they find may originate from altered water vapor concentrations.

3. How much will the temperature increase?

The Manabe-Wetherald model yields a surface temperature increase of 2.93 K for the global average, with about 2 K at low latitudes and 7–10 K near the Poles. Lacking such a model ourselves, we take another approach in estimating the temperature change. As is well known, the sun heats the air essentially through latent heat transfer. The sea at low latitudes reaches a limiting temperature based on the balance between radiative input energy and evaporative loss, which is close to 303 K (Newell *et al.*, 1978). This extreme water temperature may then be regarded as the upper limit on atmospheric temperature for air over the ocean. How much does the sea temperature change if the energy available at the surface increases by 1 W m^{-2}? The surface energy balance (for a wind speed of 3 m s^{-1}) appears in Fig. 1.

This graph is that used to derive the limit of 303 K mentioned above and is based simply on sensible and latent heat loss equations by Budyko (1974) and on a radiative transfer equation for the surface used by Privett (1960). We do not intend to discuss these equations here in detail, for they have been aired in various forms quite extensively in the literature; they are listed here for completeness:

$$Q_{LH} = 6.08(q_s - q_a)W$$

$$Q_{SH} = 2.51(T_s - T_a)W$$

$$Q_b = 0.94[\sigma T_s^4(0.56\text{–}0.08\sqrt{e_a})].$$

Here Q_{LH} is latent heat loss, Q_{SH} sensible heat loss and Q_b effective back radiation, or the difference in long-wave radiation emitted by the sea and that coming from the atmosphere. q_s is the saturation specific humidity (g kg^{-1}) at the temperature of the surface T_s, q_a the specific humidity (g kg^{-1}) of the air at temperature T_a, W the wind speed (m s^{-1}), e_a the vapor pressure (mb) of the air and σ the Stefan-Boltzmann constant. With the units given above, the Q's are in W m^{-2}. Two values of W, 3 and 6 m s^{-1}, are used in the equation for Q_{LH} for illustrative purposes, but the Q_{SH} term and the total is only shown for the lower speed. For T_a and q_a fixed (here at 300 K and 15 g kg^{-1}, respectively) the fluxes are simply a function of T_s.

The main point here is the variation of tropical sea temperature with available energy. We may treat the increased infrared energy available at the surface as additional incoming energy and use the slope of this curve to estimate the sea temperature change. The slope is $\sim$0.03 K(W m^{-2})$^{-1}$, so additional energy of $\sim$1 W m^{-2} yields a temperature increase of about 0.03 K, which we may apply to the atmosphere also. While this approach is very simple, it does yield maximum sea and air temperatures close to those observed without any unreasonable physical assumptions. The deduced change is substantially less than the 2 K found by Manabe and Wetherald for tropical latitudes. In fact, it is less than the limit of detection.

It is possible to attempt a calibration of this crude model with the observations recorded after the Mt.

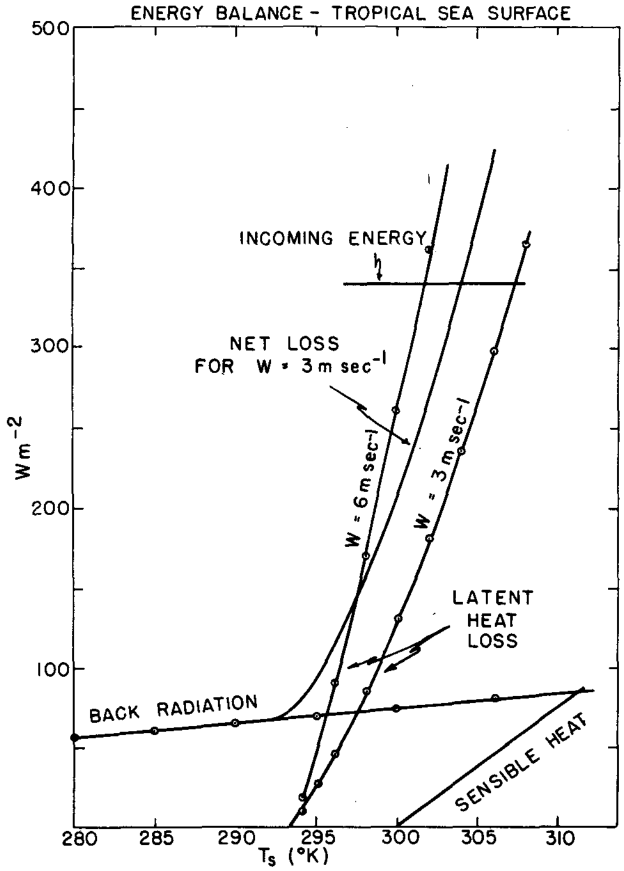

FIG. 1. The main terms in the energy balance at the tropical sea surface (for $T_a=300$ K and $q_a=15$ g kg^{-1}).

Agung eruption in 1963. Direct solar radiation received at the ground decreased while diffuse scattered radiation increased; Dyer and Hicks (1965) suggest that the total radiation values showed a significant reduction. Lamb (1970) proposes that this reduction was $\sim$7%, which to us seems to represent an upper limit based on the curves presented by Dyer and Hicks. This energy change of about 23 W m^{-2} would correspond on our model to a decrease in tropical sea surface temperature of $\sim$0.7 K. The tropical tropospheric free-air temperature was observed to decrease by $\sim$0.5 K (Newell and Weare, 1976). Paltridge (1974) has estimated a change of 0.35 K for an additional 1% of solar radiation for the atmospheric system; the corresponding average increase in solar radiation at the surface is about 3.4 W m^{-2}, yielding a sensitivity of about 0.1 K (W m^{-2})$^{-1}$ for the entire globe.

The fact that water vapor dominates CO_2 in the radiation budget has been known and discussed for many years (see, e.g., Kondratiev and Niilisk, 1960; Möller, 1963; Zdunkowski *et al.*, 1975) but it seems important to reemphasize when so much attention is being paid to CO_2.

The sensitivity deduced by Möller from radiation considerations is about 0.5 K (W m^{-2})$^{-1}$. The difference between his work and ours is evident from the slopes of the curves in Fig. 1. If the additional radiation at the surface from increased CO_2 is balanced by additional longwave radiation emitted from the surface, the surface temperature must rise following the "back radiation" curve. If the additional energy is balanced by latent heat liberation, the associated temperature rise is evidently much smaller. Kondratiev stressed that the non-radiative components of the heat budget should be included in these considerations and Möller specifically recommended inclusion of the latent and sensible heat terms as was done in Fig. 1. In fact, of course, the Manabe-Weatherald model includes these terms. Yet their final sensitivity, from a global mean surface temperature change of 3 K for additional infrared energy at the surface of $\sim$3.5 W m^{-2}, is close to 1 K (W m^{-2})$^{-1}$. Is the difference found due to the fact that we are making a comparison for the tropics, whereas their numbers apply to the globe? Or is it due to the fact that our approach does not include water vapor feedback? As a simple test of this feedback we suppose that the additional energy from CO_2 ($\sim$1 W m^{-2}) goes into the evaporation of water. Let us consider 10 days worth of evaporation corresponding to the mean residence time of water in the atmosphere. The increase in column water content is $\sim$0.3 kg m^{-2} or $\sim$1% of the content in the tropics. We recomputed the radiation profile for clear and cloudy conditions at the equator with the result that the surface loss of infrared radiation is reduced by $\sim$1 W m^{-2}. The total change after this first iteration, as it could be termed, is then 2 W m^{-2}—not far from the Manabe-Wetherald finding of 3.5 W m^{-2}. We are thus returned to the earlier question about their high sensitivity. Part of the answer may be in the fact that their surface temperature is extremely high: initially 306–307 K and $\sim$310 K after the CO_2 doubling according to their Fig. 5. Such a temperature could not be produced for the ocean according to Fig. 1 and is not in fact observed over the sea. The very high water vapor content that accompanies such temperatures at low latitudes (24–29 g kg^{-1} for 75% relative humidity) may be a prime factor in governing the changes in their model. Rowntree and Walker (1978) have also commented on the high temperature predicted by the model both before and after doubling of the CO_2.

It is assumed here that the dominant factor in controlling tropical air temperature is latent heat liberation. The air also gains energy from the land; from energy balance studies by Vernekar (1975) or the climatic maps of Budyko (1974) it can be seen that in the desert regions such as the Sahara and Australia the sensible heat energy supply is dominant. For these regions we may adopt the sensitivity of Möller quoted above, *viz.*, 0.5 K (W m^{-2})$^{-1}$. The regions occupy at most 8% of the tropics and we can assume that the additional CO_2

will provide a temperature increase of 1 K there for 2 W m^{-2} additional energy. Then weighting the changes by the areas of the desert and the ocean the net change for the tropics increases to 0.13 K, still below the limit of detection. Even using the Manabe-Wetherald values of 3.5 W m^{-2} with Fig. 1 (assuming T_a and q_a changes are small so the same curves may be used) we derive for tropical land and sea combined a temperature increase of only 0.24 K, still considerably less than the value they find of over 2 K, but now probably detectable. The value of 3.5 W m^{-2} is a mean over the globe; their values for the tropics are not published.

The conclusion is that at low latitudes the influence of doubling CO_2 on surface air temperature is less than 0.25 K, smaller by a factor of 8 than the findings generally accepted. Our finding is comparable to that by Zdunkowski *et al.* (1975). Our next step is to try to devise an approach similar to that used in Fig. 1 for the higher latitudes.

It is not our intention to diminish the importance of the anthropogenic CO_2 problem. In fact, if such different conclusions can be reached by separate paths, it becomes even more important to study the problem in greater depth.

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