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Authors
Kuhn, WR
Kraemer, D
Cicerone, RJ
et al.

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RADIATIVE EQUILIBRIUM TEMPERATURES IN THE STRATOSPHERE AND MESOSPHERE - A COMPARISON FOR THE STELLAR OCCULTATION AND BUV OZONE DATA

W. R. Kuhn, D. Kraemer, R. J. Cicerone and S. Liu

Department of Atmospheric and Oceanic Science, University of Michigan, Ann Arbor, Michigan 48109

Abstract. Recently published ozone densities in the equatorial stratosphere and mesosphere from satellite stellar occultation measurements are appreciably larger than previous independent midlatitude measurements and ozone densities predicted by photochemical theory. We have calculated radiative equilibrium temperature profiles with a stratosphere-mesosphere optimized model for each of these ozone profiles and find that the stellar occultation studies cannot be discounted on energy balance grounds.

Introduction

Recent stellar occultation measurements of nighttime equatorial ozone between 45 km and 110 km by Riegler et al. [1976, 1977] are two to three times higher than the midlatitude ozone concentrations of Krueger and Minzner [1976]. Theoretical photochemical calculations [Riegler et al., 1977] also yield ozone concentrations about three times lower than the stellar occultation results. Extra ozone sources or a very dry mesosphere (0.1 ppm H_2O) is required to fit the ozone data theoretically. Since the solar radiation absorption by ozone is the major heat source in the mesosphere and upper stratosphere we decided to calculate the temperature with a one-dimensional radiative convective model to determine how closely radiation controls the temperature structure and if one or the other O_3 profiles could be ruled out. We have compared this temperature with the temperature calculated using BUV ozone profiles [Krueger and Minzner, 1976] and with measured temperatures. The effects on the temperature of a very dry (0.1 ppm H_2O) and very wet (10 ppm H_2O) stratosphere have also been investigated.

Methodology

Radiative equilibrium temperature profiles were calculated for the stratosphere through the mid-mesosphere extending to 0.01 mb (~80 km). The radiation entering the stratosphere from below was calculated at 150 mb. The troposphere was one of fixed lapse rate given by the tropical model of McClatchey et al. [1971] although the tropospheric temperature was allowed to vary. The temperature at the upper boundary was fixed at 184°K.

The conditions imposed on the model were a stratosphere and mesosphere in radiative equilibrium, and an outgoing planetary flux equal to 2.6×10^5 ergs/cm²sec, in agreement with vonderHaar and Suomi [1971]. A radiative equilibrium stratosphere and mesosphere were first calculated with a boundary flux given by the tropical atmosphere tabulated by McClatchey et al. [1971]. If the outgoing planetary radiation did not agree with the measured radiation field, the tropospheric profile was shifted, a new boundary flux determined, and the calculations for the stratosphere and mesosphere repeated. This procedure was continued until the outgoing flux agreed to within $\pm$ 1% of that measured by vonderHaar and Suomi [1971].

The tropopsheric water vapor is from McClatchey et al. [1971]. The CO_2 and O_3 concentrations were fixed at 4.56×10^{-4} and 4.8×10^{-9} gm/gm respectively. The cloud cover was assumed to be 50% with the cloud tops located at 525 mb. Our "standard" upper air model for ozone (Figure 1) is from Krueger and Minzner [1976]. This model applies to latitudes of 30° to 60°; however, the equatorial ozone data presented earlier by Krueger [1973] yield essentially the same profile as given in their 30° to 60° latitude model. Also shown are the equatorial ozone data for the stellar occultation measurements of Riegler et al. [1977]; the water vapor mixing ratio is from Liu et al. [1976], and the CO_2 amount is 4.56×10^{-4} gm/gm.

The stratosphere ozone concentrations shown in Figure 1 are the daytime values. Since the measured equatorial values of Riegler et al. [1977] are nighttime values, we have used the calculated ratio of day-night O_3 from a one-dimensional diurnal model to obtain the daytime O_3.

The solar heating rate was calculated for a declination of 0°, and latitude 15°. A numerical integration of the heating rate (Gauss-Legendre quadrature) over hour angle was performed for ozone while an effective zenith angle of 57° was assumed for the solar heating by the near infrared bands of CO_2 and H_2O. A diffusivity factor of 1.667 was used for the terrestrial radiation field.

Most of the radiative equilibrium calculations to date have been concerned with the troposphere and extend into the mid-stratosphere. Since we are interested in temperature up to the mid-mesosphere region, we have placed the upper boundary of our model at 0.01 mb (~80 km) so that reliable results can be obtained to approximately 70 km. Since flux divergences are very small at these elevations, accurate transmission functions are required. Rather than using an average absorption for each band, we have divided the planetary spectrum into 31 spectral intervals extending to 2450 cm^{-1}. The spectral data, from McClatchey et al. [1973], contained the CO_2, H_2O, and O_3 lines. Absorption of solar radiation by CO_2 and H_2O from 500 to 12000 cm^{-1} was included with data also from McClatchey et al. [1973]; thirty spectral intervals were used. The continuum absorption by water vapor which is important in the low latitude troposphere was also included; we used the absorption coefficients from Bignell [1970] for the spectral region extending out to 1250 cm^{-1}. The absorption cross sections for ozone and oxygen and the photon fluxes were from Ackerman [1971].

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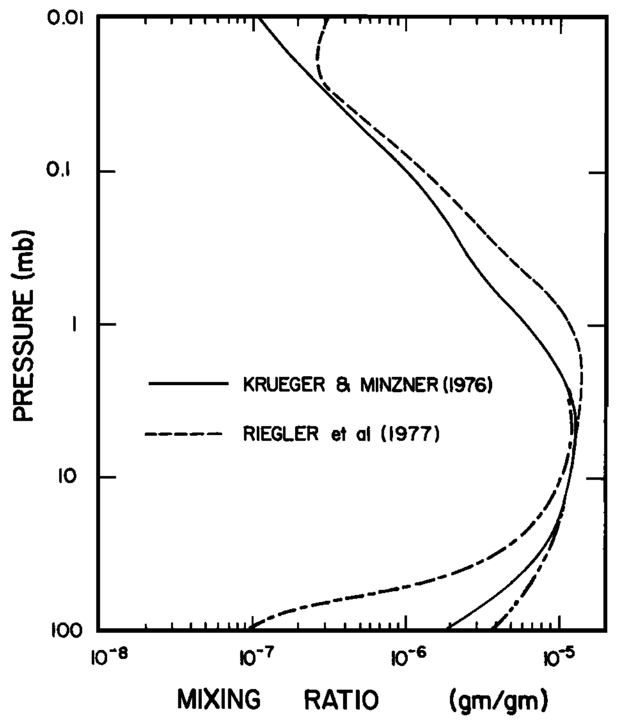

Figure 1. Ozone mass mixing ratio profiles from data of Krueger and Minzner [1976] and Riegler et al. [1977] modified for daytime conditions. The profile — - — corresponds to 60°N [Dütsch, 1974] while — -- — is an equatorial profile [McClatchey et al., 1971].

The transmission functions for CO_2, H_2O, and the 9.6 µm band of O_3 were calculated from the spectral data previously referenced. The lines were grouped into the various spectral intervals. Average line intensities for five intensity decades within each interval were then determined for temperatures ranging from 325°K to 150°K at 25°K intervals. The line strengths used in the calculations were found by interpolating on this table with the Curtis-Godson temperature corresponding to that particular mass path. The Voigt profile was used for all calculations so that a matching of the Doppler and Lorentz approximation in the overlap region, which occurs in the mid-stratosphere, was not required. Comparisons to experimental data from Palmer [1960] and Burch et al. [1962], (for H_2O), and Burch et al. [1962], (for CO_2), and Bartman et al. [1976] and McCaa and Shaw [1967], (for O_3) showed that the computed transmissivities were within the range of uncertainty of the laboratory measurements.

The radiative equilibrium temperature was determined by a simple time-marching scheme first applied by Manabe and Möller [1961]. An initial guessed temperature is corrected by adding the calculated radiative temperature change to the previous temperature and continuing this procedure until the heating or cooling rate is less than some prescribed value, in our case 0.01 deg/day. We find this approximate radiative equilibrium is reached (including the correct outward planetary radiation) after approximately 120 iterations.

The sensitivity of the model to the boundary conditions and the pressure grid have been checked. A variation of the temperature by 10°K at the upper boundary (0.01 mb) causes the temperature at the 0.1 mb level to vary by less than 1°K. A variation in cloud height from 825 to 375 mb causes the lower stratospheric (up to 70 mb) temperature to change by about 5°K. Indeed, above 30 mb, the troposphere influences the temperature by less than a few degrees regardless of the tropospheric data used.

Solar absorption by the near infrared bands of CO_2 and H_2O do influence the radiative equilibrium temperature. If these bands are excluded then the temperature is some 3° to 6°K lower throughout the stratosphere-mesosphere.

Accurate modeling of the transmission of the 9.6 µm O_3 band is necessary if temperatures near the stratopause are to be calculated. If cooling by this band is neglected, we find the temperature to be 20°K higher at the stratopause. In the low stratosphere the temperature would be about 4°K less because the 9.6 µm band produces a heating rather than cooling in the vicinity of the tropopause.

Results

The calculated temperature profiles along with those from the U.S. Standard Atmosphere Supplements are given in Figure 2. Ideally, one would like to compare the radiative equilibrium temperature with a temperature observed simultaneously with the ozone measurements; however, this was not possible since rocketsonde observations at the same time and location were not available.

The most meaningful comparison is for the equatorial case, where the Riegler et al. [1976] data were obtained, but we have also included calculations for the latitudes of 45° and 60°. For pressures less than about 5 mb (~35 km) we assumed the same ozone concentrations as in the equatorial calculations (see Figure 1). For lower elevations we used the mid-latitude ozone concentrations from McClatchey et al. [1971] and the polar concentrations from Dütsch [1974]. The tropospheric boundary flux was calculated for temperature and composition appropriate to the particular latitude.

The largest difference in temperature for the equatorial case between the calculated profile based on the Riegler et al. [1976] ozone data and the observed temperature is about 16°K and occurs near the stratopause and in the lower mesosphere. The Krueger and Minzner [1976] data yield the largest difference, about 7°K, with the observed temperature near the stratopause. The calculated temperatures from the two sets of ozone differ most in the low mesosphere where the Riegler et al. [1976] data give a temperature some 16°K higher than the ozone data from Krueger and Minzner [1976].

The stratosphere is very close to radiative equilibrium and the calculated temperatures for the two different ozone profiles differ by at most 5°K. The Riegler et al. [1976] data give a slightly lower temperature than the Krueger and Minzner [1976] ozone data even though the latter corresponds to the smaller ozone concentrations; this effect is due to greater cooling by the 9.6 µm O_3 band for the Riegler et al. [1976] data as well as the solar heating occurring higher in the atmosphere. At the stratopause and in the mesosphere, the Riegler et al. [1976] ozone data

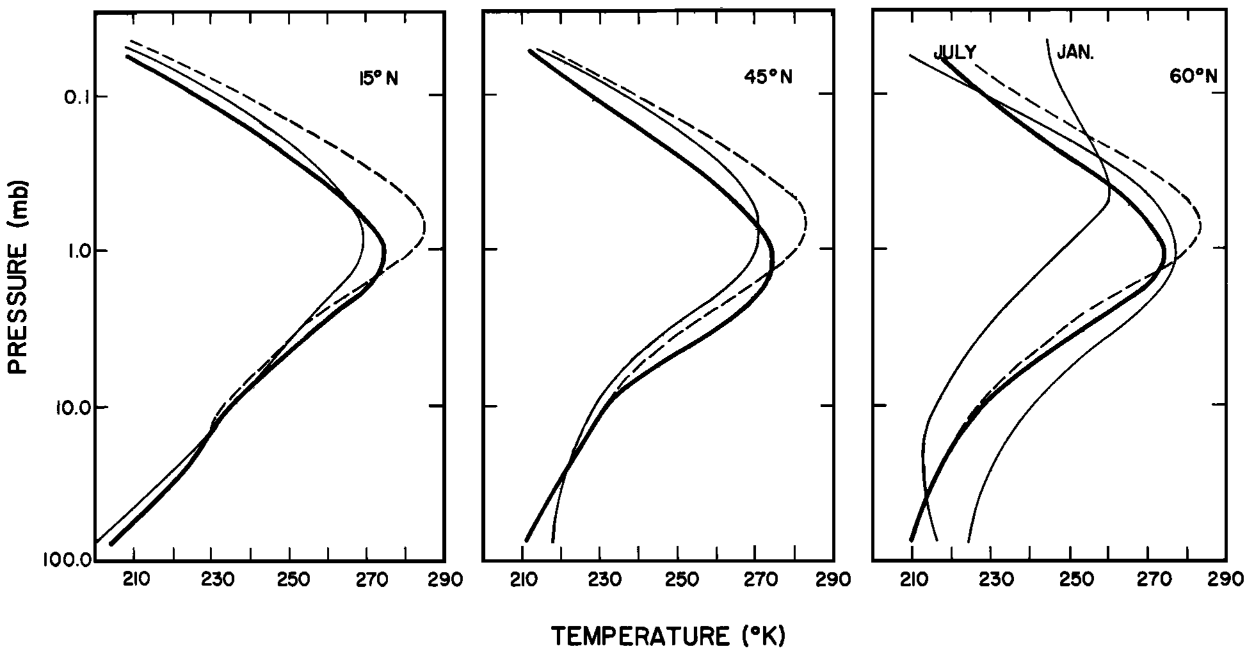

Figure 2. Radiative equilibrium temperatures (heavy line) for latitudes 15°N (equinox), 45°N (equinox) and 60°N (equinox). The thin line corresponds to the temperature from the U.S. Standard Atmosphere Supplements, 1966. The dashed line refers to the radiative equilibrium temperatures for the Riegler et al. [1977] ozone data, while the heavy line is for the Krueger and Minzner [1976] ozone data.

yield the higher temperatures because the larger amount of ozone produces a greater solar heating. A cursory examination of the equatorial calculations shows that the Krueger and Minzner [1976] data produce a radiative equilibrium temperature nearer to that of the mean 15° latitude temperature. Nevertheless, because of uncertainties in water vapor concentrations, heat transport, and our lack of knowledge of the actual temperature at the time of the measured ozone it would be difficult to conclude that the Riegler et al. [1976] data are erroneous.

If the Riegler et al. [1976] data would apply to mid and high latitudes, then the radiative equilibrium temperatures would be as shown in Figure 2. Throughout much of the midlatitude stratosphere and in the upper mesosphere, it appears that the Riegler et al. [1976] ozone data would yield a temperature profile nearer the observed temperature than the Krueger and Minzner [1976] data. Again because of the similarities of the calculated temperature distributions, it would be difficult to argue that the Riegler et al. [1976] ozone data are erroneously high based on the radiative equilibrium temperature considerations.

The thermal balance requirements for the equatorial case are shown in Table 1. The energy which must be removed by non-radiative processes to maintain the observed temperature is given in ly day^{-1} and represents an atmospheric column extending from the top of the atmosphere down to the indicated pressure level. Note that the Krueger and Minzner [1976] ozone data require that the mesosphere be a weak energy sink while according to the Riegler et al. [1976] data the mesosphere must lose energy. Non radiative processes must remove energy from the stratosphere for both sets of ozone data. The stratospheric thermal excess for the Riegler et al. [1976] data is actually slightly less than from the Krueger and Minzner [1976] data since the larger amount of ozone in the former case is causing a larger cooling from the 9.6 μm band. Although there is an increase in solar heating from the larger ozone concentrations, this heating occurs at the higher, i.e., mesospheric elevations. For example, the thermal excess is 14% less in the interval from .875 to 50 mb for the Riegler et al. [1976] data than for the Krueger and Minzner [1976] data. Thus, energy balance requirements do not allow one to negate either of the ozone data sets.

Table 1. Non-radiative energy requirements (ly day^{-1}) necessary to produce the observed U.S. Standard Atmosphere 15° equinox temperature profile for the above referenced O_3 models. Energy is for an atmospheric column extending from 0.01 mb down to the indicated pressure level.

Pressure (mb)	thermal excess (ly day^{-1}) Riegler et al.	Krueger and Minzner
0.15	-0.03	-0.05
0.425	0.15	-0.11
0.875	0.79	-0.11
1.75	1.34	0.35
4.25	1.59	1.07
8.75	1.70	1.42
15.0	1.70	1.51
47.0	5.92	5.88
103.0	10.58	10.53

Since there is much uncertainty in the stratosphere-mesosphere water vapor distributions, we determined its influence on the temperature profile for the cases with constant mixing ratios of 10^{-7} and 10^{-5} gm/gm. The actual water vapor distribution is probably well within these limits. The largest variation in radiative equilibrium temperature occurs in the low mesosphere where the difference is about 15°K. Thus, it is doubtful that the actual water vapor distribution will appreciably change the calculated thermal profiles.

It is clear that calculations of latitudinal distributions of infrared cooling and solar heating rates which are necessary for upper air general circulation models will only be very approximate until upper stratosphere-mesosphere ozone data are available. The Krueger and Minzner [1976] and Riegler et al. [1976] ozone data yield net equatorial heating rates whose differences are as large as 5°K day^{-1} in the lower mesosphere. The radiative equilibrium temperatures differ as much as 15°K.

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