GCA-TR-67-18-N

RADIATIVE TRANSFER AND THE THERMAL STRUCTURE OF PLANETARY ATMOSPHERES

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TECHNICAL REPORT

Contract No. NASW-1574

November 1967

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GCA TECHNOLOGY DIVISION
Bedford, Massachusetts

Prepared for

NATIONAL AERONAUTICS AND SPACE ADMINISTRATION
HEADQUARTERS
Washington, D. C.

FOREWORD

This paper was presented as an invited review paper at the Survey Symposium on Radiation at the XIV General Assembly of the International Union of Geodesy and Geophysics, September 25 - October 7, 1967.

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ABSTRACT

The mean surface temperature and mean vertical distribution of temperature of a planetary atmosphere are, to a large extent, determined by radiative processes. Given the basic physical characteristics of the planetary atmosphere, such as composition, surface pressure, and albedo, it is possible, through the application of radiative transfer theory, to compute the temperature structure of the planetary atmosphere. For the Earth's atmosphere the problem has been to determine theoretically the mean vertical temperature profile from the surface up to the thermosphere. An additional problem, related to climatic change, is the effect of changes in the basic physical characteristics, such as composition or cloudiness, on surface and atmospheric temperatures. Applications of radiative transfer theory to the atmospheres of the other planets have been aimed at deriving estimates of surface and atmospheric temperatures on Mars and estimates of the greenhouse effect on Venus. The theoretical models and results of recent studies of these topics are reviewed in the present survey paper.

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1. INTRODUCTION

To a large extent, the mean surface temperature and vertical distribution of temperature in a planetary atmosphere are controlled by radiative processes. Convection and condensation-evaporation processes may modify the thermal structure that would result from radiation alone. In this review paper, we shall be concerned mainly with the problem of theoretically estimating the mean thermal structure of planetary atmospheres. The theoretical models that have been developed to solve this problem depend primarily on radiative transfer considerations, but many of them also include the effect, if not the detailed mechanisms, of convection and condensation-evaporation.

We shall discuss problems associated with the planets Earth, Mars, and Venus. For the Earth, the basic problem is to derive from first principles the observed mean temperature profile from the surface up to the thermosphere and to predict the effects of changes in atmospheric composition on the thermal structure. For the planets Mars and Venus, the basic problem is to predict the mean surface temperature and vertical temperature profile on the basis of available information on atmospheric physical properties such as composition, surface pressure, and albedo. By means of a selective review, we shall summarize the results of recent research on these problems.

As background for a discussion of the mean thermal state of each of the planets and comparisons between the planets, it is instructive to review the radiative equilibrium temperatures that these planets would have in the absence of an atmosphere. The radiative equilibrium temperature of a planet is computed from an equation representing a balance between the planet's blackbody emission and the net incoming solar radiation, which depends upon the solar constant, the planet's distance from the sun and its albedo.

$$\sigma T_e^4 = \frac{S.C.}{4R^2} (1 - A) \tag{1}$$

where σ is the Stefan-Boltzmann constant, T_e is the planet's radiative equilibrium or mean temperature in the absence of an atmosphere, which we shall call the effective temperature of the net incoming solar radiation, S.C. is the solar constant, R is the planet's distance from the sun in astronomical units, and A is the planet's albedo.

Figure 1 shows the effective temperature of Mars, Venus, and Earth for a range of possible albedos for these planets. In the absence of atmospheres, Venus, though closer to the Sun than Earth would have a mean temperature similar to Earth's because of its greater albedo. Mars, being further from the Sun than Earth and with an albedo only slightly less than Earth's, would have a lower temperature than Earth.

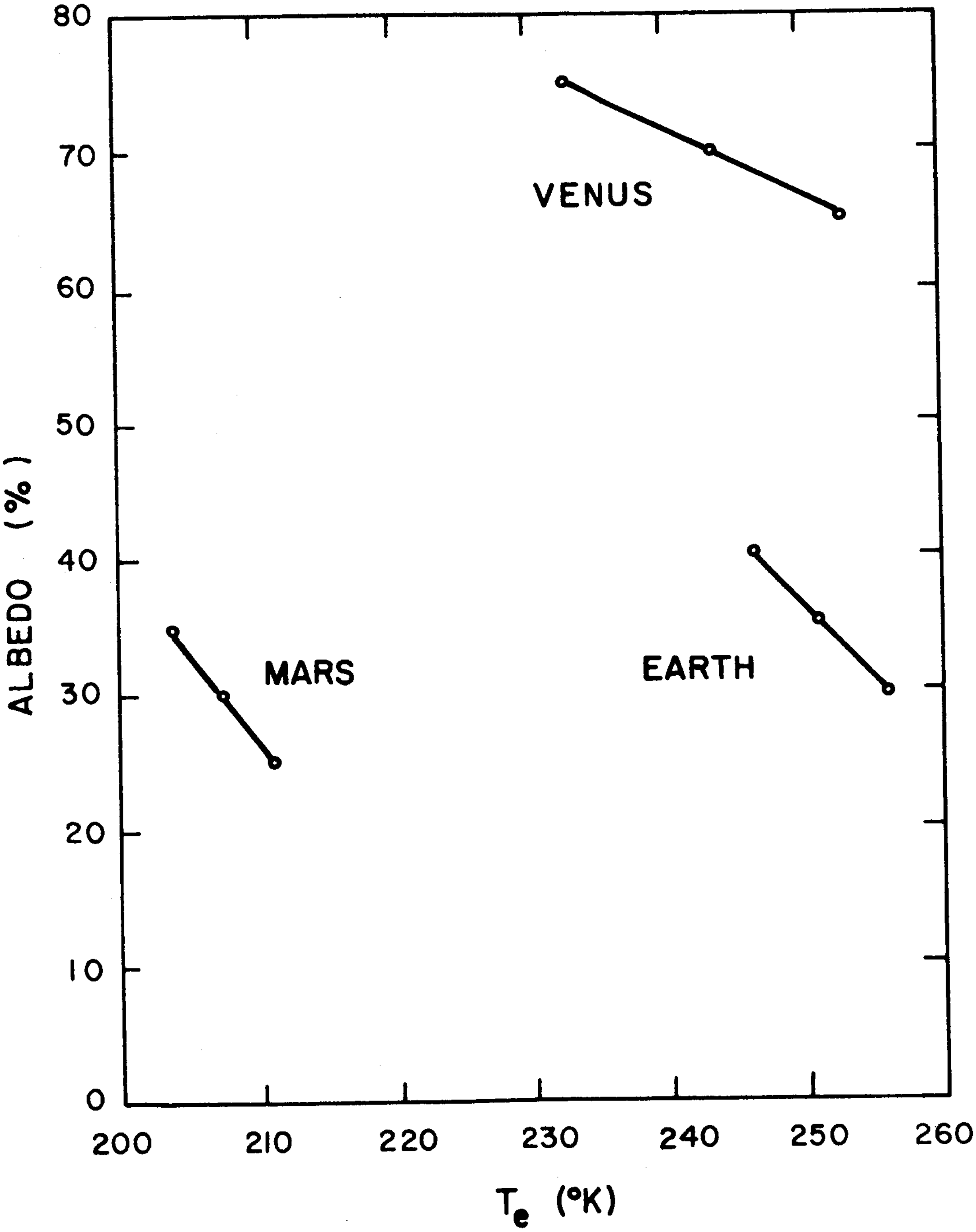

Figure 1. Radiative equilibrium temperatures of Earth, Mars, and Venus.

2. EARTH

In connection with the Earth's atmosphere, one of the basic problems is to derive theoretically from first principles the observed mean vertical temperature distribution from the surface to the upper atmosphere. That is, given the planet's physical characteristics, the composition of its atmosphere, etc., can one derive the mean vertical temperature profile. Although the problem has not been completely solved, significant advances have been made over the past few years.

In a series of papers (Manabe and Möller, 1961; Möller and Manabe, 1961; Manabe and Strickler, 1964; and Manabe and Wetherald, 1967), Manabe and Möller and their colleagues have applied radiative transfer theory to the problem of determining the thermal state of the Earth and its atmosphere. In their most recent papers, they have used a thermal equilibrium model for the calculations. The requirements for thermal equilibrium are:

(1) At the Top of the Atmosphere — A balance between net incoming solar radiation and outgoing long-wave radiation.

(2) Within the Atmosphere — For those layers in which radiative energy exchange leads to a radiative equilibrium temperature lapse rate that is less than a prescribed critical convective lapse rate, local radiative equilibrium is assumed. For those layers in which radiative energy exchange would lead to a temperature lapse rate greater than the prescribed critical convective lapse rate, a balance between the radiative energy loss and the amount of convective energy required to maintain the critical convective lapse rate is assumed.

(3) At the Surface — A balance between the net radiative energy gain and the convective energy loss. The convective energy loss is equal to the amount of heat that must be transferred to the atmosphere to prevent the formation of a super-convective lapse rate. Thus, the thermal equilibrium model includes the effect of convective processes, which is to eliminate unstable layers, without dealing with the actual dynamics of convection.

Given a vertical distribution of water vapor, carbon dioxide, ozone, and cloudiness — that is, the atmospheric absorbers of radiation — the model can be used to compute the temperature of the Earth's surface and the vertical temperature profile. The details of the transmission models for the various absorption bands, the transmission, reflection, and absorption properties of the clouds, and the surface albedo model are presented in the original papers.

Figure 2 shows a comparison of the vertical temperature profiles computed for three cases: pure radiative equilibrium; thermal equilibrium with a critical convective lapse rate equal to the adiabatic lapse rate; and thermal equilibrium with a critical convective lapse rate equal to the observed mean tropospheric lapse rate of 6.5°C/km, which is close to the average moist adiabatic lapse rate in the troposphere. All three are for average conditions of insolation and absorber distribution and for clear skies. The pure radiative

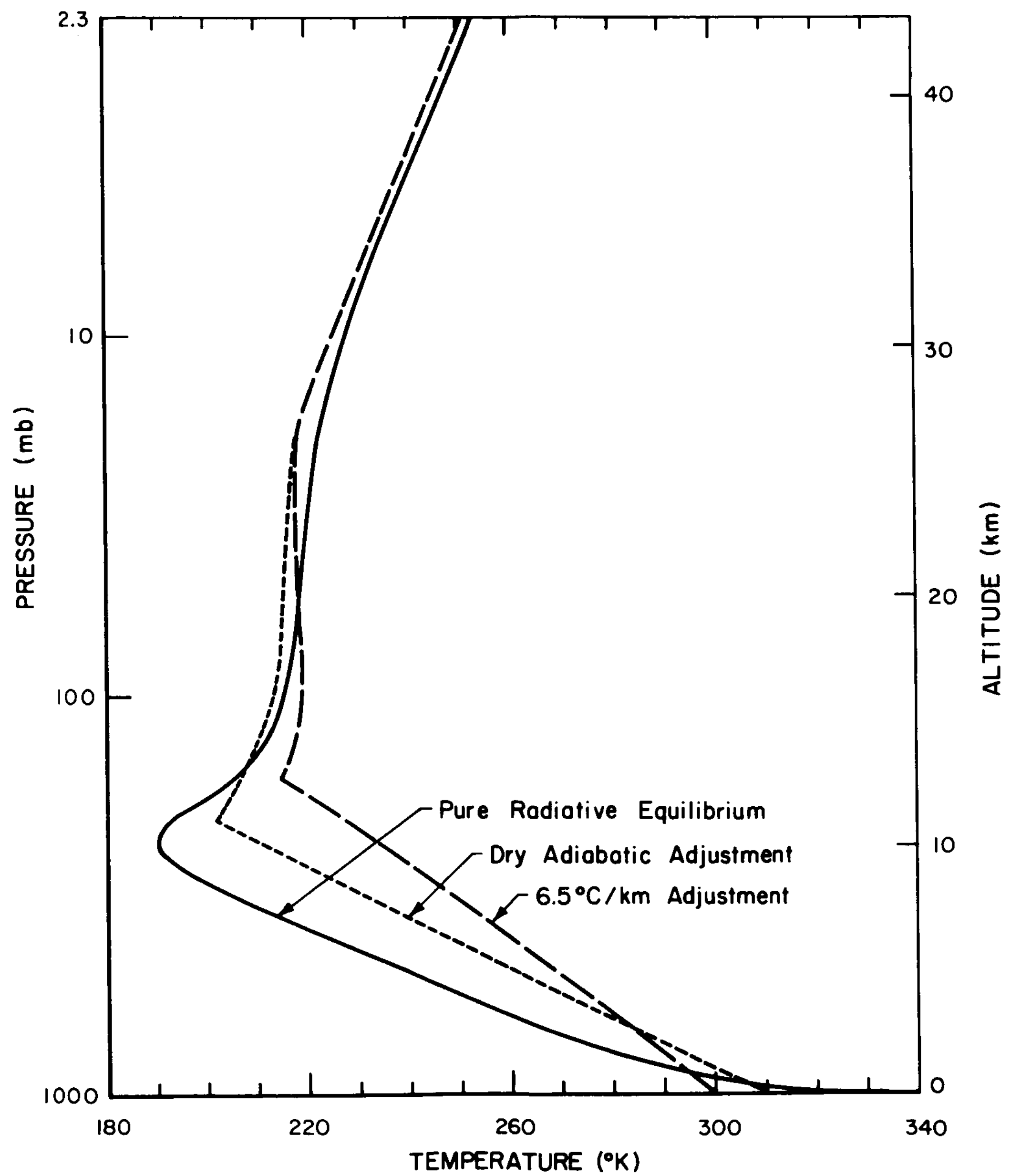

Figure 2. Pure radiative equilibrium and thermal equilibrium temperature profiles for Earth. (After Manabe and Strickler, 1964)

equilibrium calculation leads to a surface temperature of 332°K and a super-adiabatic tropospheric layer. The effect of convection is to lower the surface temperature and raise the tropospheric temperatures. With a prescribed convective lapse rate of 6.5°C/km, the computed surface temperature is 300°K, which is closer to the observed mean of 288°K than in the pure radiative equilibrium case, but still somewhat too high.

The next step is to include the effects of clouds. Figure 3 shows a comparison of the thermal equilibrium temperature profiles computed with and without average cloudiness and the U.S. Standard Atmosphere temperature profile. The computed surface temperature for the case of average cloudiness is 287°K, in excellent agreement with the observed mean surface temperature. The computed vertical temperature profile is also in good agreement with the observed mean temperature profile as represented by the U.S. Standard Atmosphere. That a calculation such as this can reproduce the observed mean temperature profile with good fidelity is indeed encouraging.

A problem of importance to studies of climate change and control is the effect of variations in atmospheric composition on the mean surface temperature of the Earth. One of the questions that arises is the effect of changes in the average amount of cloudiness. Qualitatively, the effect is two-fold. On the one hand an increase in the amount of cloudiness would result in an increased albedo and, hence, tend to lower the surface temperature. On the other hand, an increase in the amount of cloudiness would result in an increased greenhouse effect, which would tend to raise the surface temperature. Ohring and Mariano (1964a) used a simple theoretical model to estimate this effect. It is assumed that

(1) The atmosphere is grey in the infrared (that is, the absorption coefficient is independent of wavelength).

(2) Clouds behave as blackbodies in the infrared.

(3) The Earth's atmosphere consists of two layers: a troposphere, in which the temperature decreases linearly with height at a given rate, and a stratosphere in which the temperature is constant with height.

Two radiative equilibrium conditions are assumed to prevail: 1) There is a balance between net incoming solar radiation and outgoing infrared radiation at the top of the atmosphere, and 2) the stratosphere is in gross radiative equilibrium — that is, the net flux of infrared radiation at the tropopause is equal to the outgoing flux of infrared radiation at the top of the atmosphere.

The input parameters are the incoming solar radiation, the amount of clouds, the infrared opacity of the clear atmosphere, the tropospheric lapse rate, and the ratio of cloud-top pressure to surface pressure. The infrared opacity of the clear atmosphere is taken to be 1.6, which corresponds to an infrared flux transmissivity of 10 percent. The tropospheric lapse rate is assumed to be 6°C/km. The cloud-tops are assumed to be located at one level in the atmosphere, which is taken to be 500 mb. Computations were performed

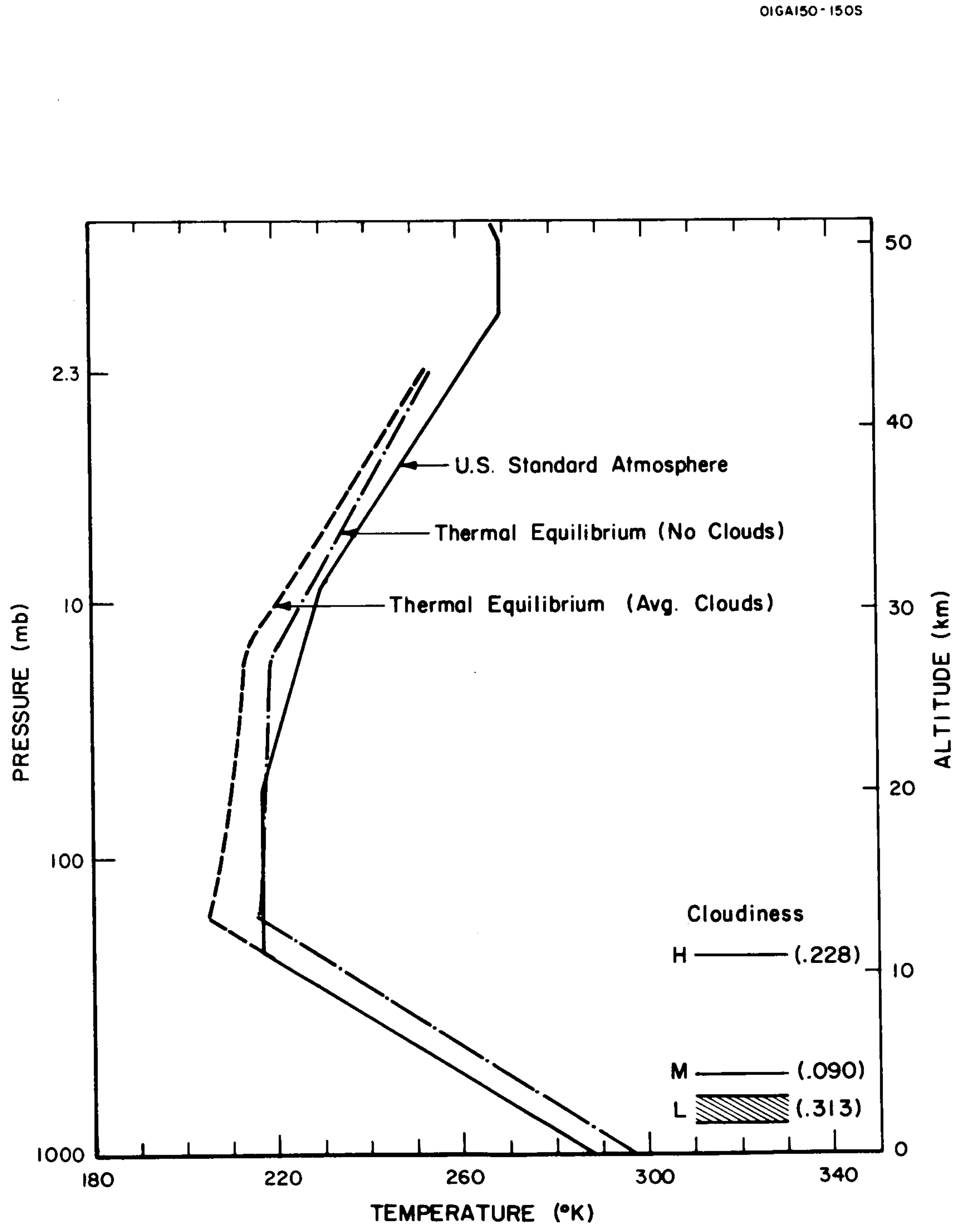

Figure 3. Thermal equilibrium temperature profiles with and without cloudiness compared to U.S. Standard Atmosphere. Amounts and heights of clouds are shown in lower right corner. (After Manabe and Strickler, 1964)

for three different amounts of total cloudiness: the average amount, 50 percent; no cloudiness at all, 0 percent; and a large amount of cloudiness, 90 percent. The planetary albedos for the different amounts of cloudiness are based upon computations by London (1957). The computed surface temperatures are shown in Table 1. The ratio T_s/T_e is the ratio of computed surface temperature to the effective temperature of the net incoming solar radiation. This ratio is a measure of the greenhouse effect. The results indicate that even with quite dramatic changes in the amount of cloudiness, the Earth's average surface temperature would change by less than 10°K. The last column of the table, (τ_t/τ_g), is the computed ratio of tropopause pressure to surface pressure for the model. For the case of 50 percent cloudiness, good agreement is obtained with the observed ratio.

TABLE 1. CHANGES IN THE AMOUNT OF CLOUDINESS AND THE AVERAGE SURFACE TEMPERATURE OF THE EARTH. (AFTER OHRING AND MARIANO, 1964)

AMOUNT OF CLOUDINESS (%)	T_s/T_e	T_e(°K)	T_s(°K)	τ_t/τ_g
0	1.13	270	305	0.263
50	1.19	252	299	0.181
90	1.23	236	290	0.131

Detailed calculations of the effect of cloudiness on surface temperature have been made with the more accurate thermal equilibrium model (Manabe and Wetherald, 1967). Figure 4 illustrates the effect of changing the amount of one type of cloud while holding the amount of other cloud types constant at their average values. Low refers to low clouds, AS to middle clouds, and CI to cirrus type high clouds. FB stands for full black, indicating that the cloud was assumed to behave as a blackbody in the infrared. HB stands for half-black. Increases in the amount of low and middle clouds lead to lower surface temperatures, while an increase in the amount of cirrus, even when cirrus is taken to be half-black, leads to higher surface temperatures. This difference is due mainly to the higher albedos of the low and middle clouds.

Another question of considerable interest is the effect of changes in the atmospheric carbon dioxide content on the Earth's surface temperature. Qualitatively, an increase in CO_2 content would increase the greenhouse effect and, hence, lead to an increase in surface temperature. Discussions and calculations of this effect have been presented by Plass (1956), Kondratiev and Niilisk (1960), Kaplan (1960), and Möller (1963). However, these calculations were based upon

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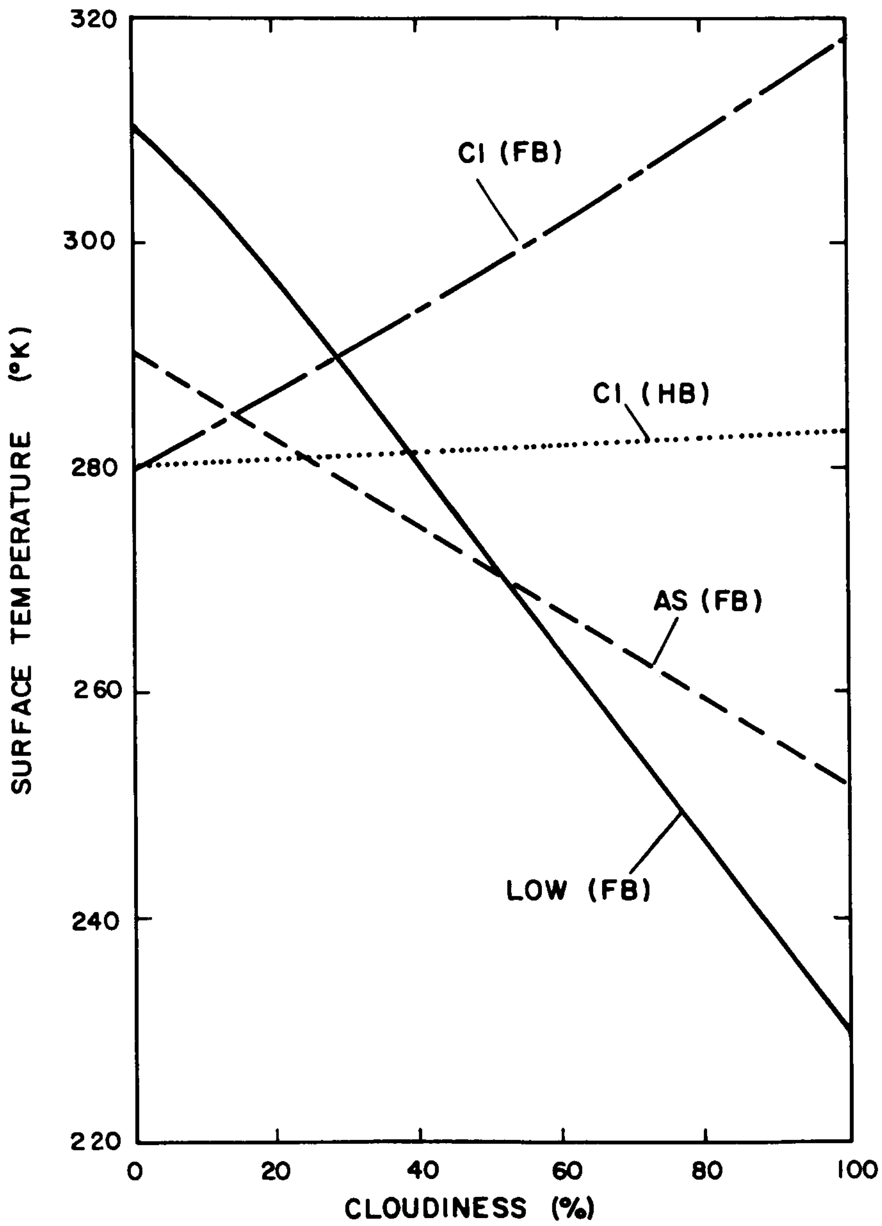

Figure 4. Earth's surface temperature as a function of cloudiness. (After Manabe and Wetherald, 1967)

considerations of the radiation balance at the Earth's surface rather than that of the entire surface-atmosphere system. Recent calculations by Manabe and Wetherald (1967) with their thermal equilibrium model are probably more reliable. Table 2 shows the effect on the surface temperature of either doubling or halving the present atmospheric carbon dioxide content. The magnitude of the change in surface temperature depends upon whether one assumes that the absolute humidity remains constant or the relative humidity remains constant as the CO_2 amount is changed. The greater change in surface temperature for the fixed relative humidity (~2°C versus ~1°C) is due to the increase in water vapor abundance associated with the increased temperature, which further enhances the greenhouse effect. Observations indicate that in the present century the atmospheric CO_2 content has increased by 10 to 15 percent due to the burning of fossil fuels. Based on the above results, the Earth's average surface temperature may have increased by 0.1 to 0.2°C in the same period.

TABLE 2. EFFECT OF CHANGES IN THE ATMOSPHERIC CO_2 CONTENT ON THE EARTH'S AVERAGE SURFACE TEMPERATURE. (AFTER MANABE AND WETHERALD, 1967)

CHANGE OF CO_2 CONTENT (ppm)	FIXED ABSOLUTE HUMIDITY		FIXED RELATIVE HUMIDITY	
	Average Cloudiness	Clear	Average Cloudiness	Clear
300 → 150	-1.25°C	-1.30°C	-2.28°C	-2.80°C
300 → 600	+1.33°C	+1.36°C	+2.36°C	+2.92°C

In their latest paper, Manabe and Wetherald (1967) also discuss the effects of changes in ozone and water vapor content on surface and atmospheric temperatures. The interested reader is referred to their original paper for the details and results of these calculations.

Of considerable interest to meteorologists is the radiative equilibrium temperature profile of the mesosphere — the region from 20 km to 80 km. Although there have been a number of previous calculations, none considered the problem as the solution of a joint radiative-photochemical equilibrium problem. That is, the radiative equilibrium temperature depends upon the ozone concentration; while the photochemical equilibrium ozone concentration depends upon the temperature. Recently Leovy (1964) solved the joint radiative-photochemical problem. The radiative balance is assumed to be between the absorption of solar radiation by molecular oxygen and ozone and infrared emission by carbon dioxide and ozone. Figure 5 shows a comparison between Leovy's computed mean radiative equilibrium temperature profile and the 1962 U.S. Standard Atmosphere, which represents the observed mean temperature profile. The computed temperature profile reproduces

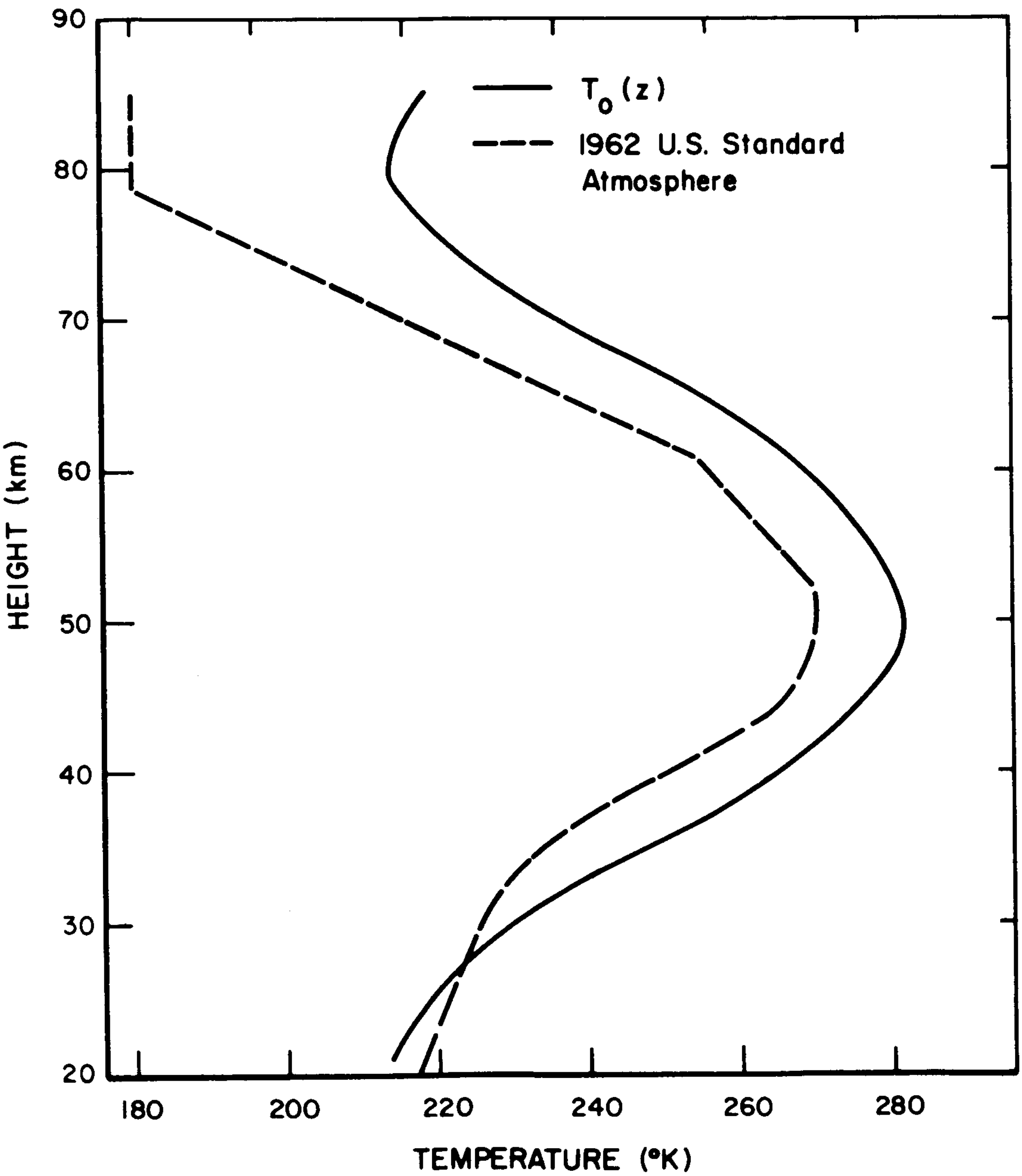

Figure 5. Radiative equilibrium temperature profile of upper atmosphere compared with 1962 U.S. Standard Atmosphere. (After Leovy, 1964)

quite well the major features of the observed mean temperature profile — that is, the temperature peak at about 50 km due to absorption of solar radiation by ozone and the temperature minimum at about 80 km. The differences between the two profiles may be ascribed to possible non-radiative energy sources and sinks, simplifying assumptions in the calculations, and even uncertainties in the observed mean temperature profile for this region of the atmosphere. Further research is required to explain these differences.

3. MARS

To obtain theoretical estimates of temperatures on Mars one must know something about the physical characteristics of the Martian atmosphere. Earth-based spectroscopic and photographic observations and the Mariner IV fly-by observations indicate that the Martian atmosphere is predominantly composed of carbon dioxide and has a surface pressure of the order of 1/100 of the Earth's surface pressure (Table 3). There are still major uncertainties in some of these parameters. For example, the surface pressure may be between 5 mb and 14 mb, and the carbon dioxide amount may be between 50 and 100 percent. The atmosphere is quite dry. Clouds apparently occur rarely and may be composed of dust, ice, or carbon dioxide.

TABLE 3. PHYSICAL PARAMETERS FOR MARTIAN ATMOSPHERE

ALBEDO	SURFACE PRESSURE	COMPOSITION	CLOUDS	$\Gamma_{ad.}$	T_e
0.3	7 mb	75% CO_2	Occasional —	4.5°/km	208°K
		25% Nitrogen/Argon	Dust, H_2O, CO_2?		
		10^{-3} cm prec. H_2O			

With such a thin atmosphere and little water vapor and cloudiness, one would expect a small greenhouse effect on Mars. This is borne out by the theoretical calculations of Ohring et al. (1962) who made an early estimate of the maximum greenhouse effect on Mars. The estimate is based upon the condition of a balance at the top of the atmosphere between net incoming solar radiation and outgoing infrared radiation. To maximize the greenhouse effect, estimated maximum amounts of carbon dioxide, water vapor, and ozone were assumed and an atmospheric lapse-rate equal to the dry-adiabatic lapse rate was adopted. The results of the calculations are shown in Table 4. The ratio T_s/T_e can be thought of as the magnitude of the greenhouse effect; T_s is the surface temperature and T_e is the temperature that Mars would have in the absence of a greenhouse effect. The results indicate that even if the greenhouse effect is maximized, Mars would have a smaller greenhouse effect than the Earth. The results of these calculations also indicated that carbon dioxide was, by far, the most important contributor to the Martian greenhouse effect.

As improved estimates of surface pressure and atmospheric composition became available, it was desirable to estimate the average greenhouse effect and surface temperature on Mars with more sophisticated models. The third column of Table 4 shows the results of calculations with a thermal equilibrium model similar to that used by Manabe and his co-workers for the Earth's atmosphere.

The computed average surface temperature is 216°K, which compares quite well with the observed average surface temperature of 220°K that is indicated by Earth-based observations of the planet's microwave emission.

TABLE 4. GREENHOUSE EFFECT ON MARS AND EARTH

	MARS (Calculated)		EARTH (Observed)
	MAXIMUM GREENHOUSE EFFECT	AVERAGE GREENHOUSE EFFECT	AVERAGE GREENHOUSE EFFECT
T_s/T_e	1.065	1.04	1.15
T_s (°K)	222	216	288
T_s (°K)	Observed	~220	

Theoretical calculations of the mean vertical temperature profile on Mars have been performed by several investigators (see, for example, Goody, 1957; Arking, 1963; Prabhakara and Hogan, 1965; Ohring et al., 1967). Figure 6 shows the results of a calculation by Prabhakara and Hogan (1965). In their model, the surface temperature is assumed and the vertical temperature profile is computed on the basis of a radiative-convective equilibrium concept, which is quite similar to the thermal equilibrium concept. Absorption of solar energy is by oxygen, ozone, and carbon dioxide, and emission of infrared radiation by carbon dioxide. Ozone and oxygen have not be detected in the Martian atmosphere, and even the small amount of oxygen assumed by Prabhakara and Hogan (70 cm-atm) should probably be considered an upper limit. For their transmission functions for carbon dioxide, they use a "statistical" model, the theoretical band parameters being based upon the transmittance tables of Stull, Wyatt, and Plass (1963). The major features of the computed Martian temperature profile, when compared to the temperature profile of the Earth's atmosphere, are the absence of a warm mesosphere and the relatively low height (< 10 km) of the tropopause. The temperature peak at 50 km in the Earth's atmosphere is due to the absorption of solar radiation by the ozone layer. Even with the upper limit values of ozone adopted by Prabhakara and Hogan there is no indication of a temperature peak above the Martian troposphere.

Since the Martian surface pressure and atmospheric composition are somewhat uncertain, it is valuable to calculate the mean surface temperature and temperature profile for a range of these parameters that covers present uncertainties.

Figure 7 shows the results of such calculations with a thermal equilibrium model (Ohring and Mariano, 1967). The adopted physical models are shown in the

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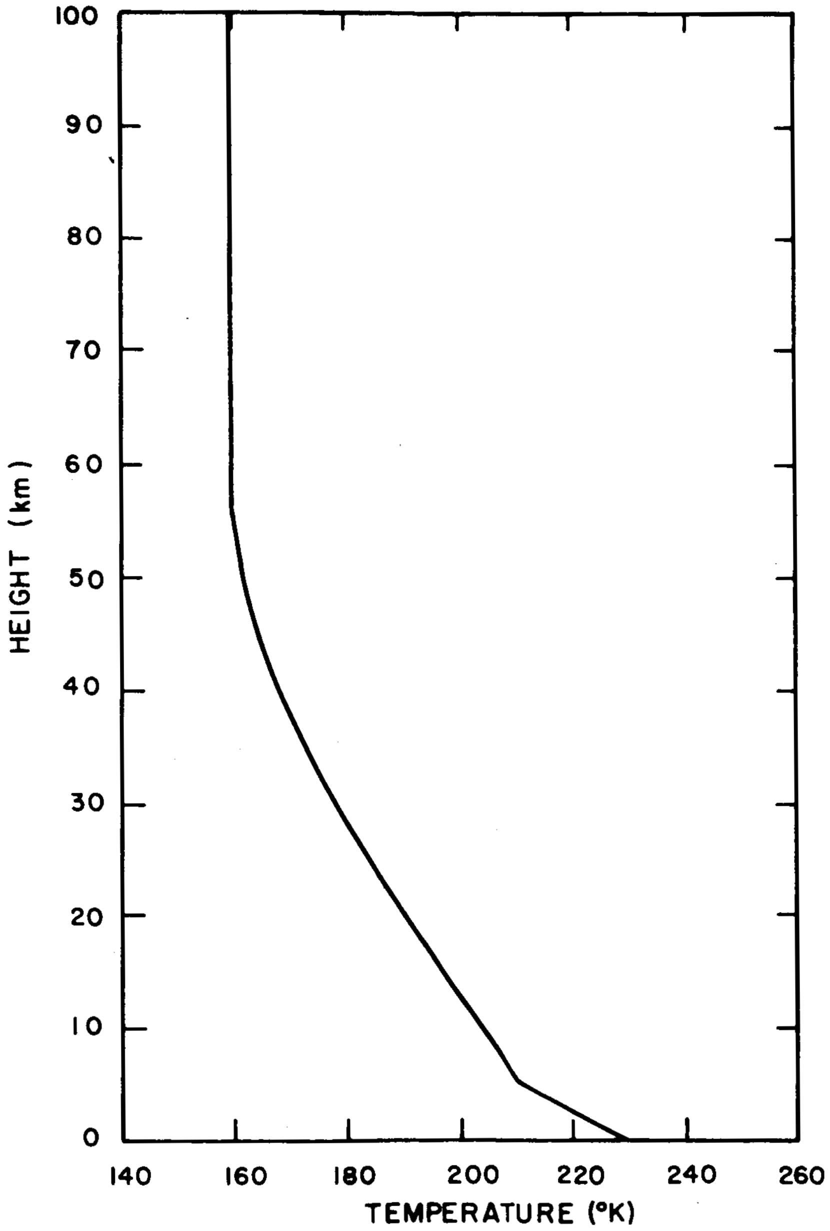

Figure 6. Calculation of vertical temperature profile for Mars. (After Prabhakara and Hogan, 1965)

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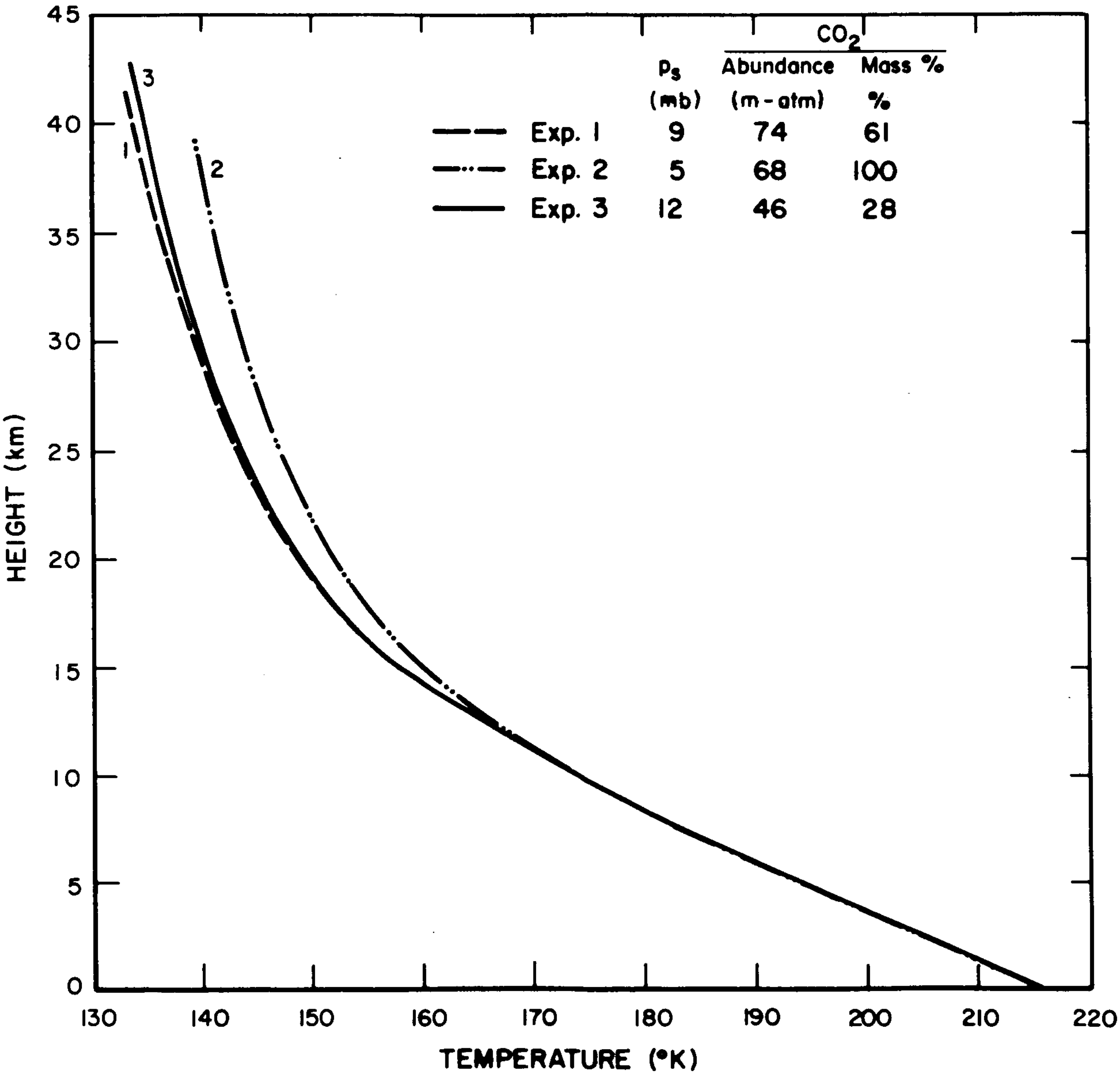

Figure 7. Thermal equilibrium temperature profiles for Mars.

graph. Absorption of solar radiation by the near infrared bands of water vapor and carbon dioxide, and emission of infrared radiation by carbon dioxide's 15μ absorption band and water vapor's rotational band are considered in the calculations. The amount of water vapor is 10^{-3} cm precipitable water. For the 15μ band transmittance and the rotational water vapor band transmittance, the model of Rodgers and Walshaw (1966) is adopted. For the absorption of near-infrared solar radiation, the absorption models of Houghton (1963) are adopted. The most interesting result of this comparison is the relative insensitivity of the computed surface and atmospheric temperatures to changes in the surface pressure and carbon dioxide content that are representative of the current uncertainties associated with these quantities. The computed surface temperatures are within one degree of each other and even at the upper levels the differences are less than 5°C. This suggests that although the surface pressure and carbon dioxide content are still somewhat uncertain, we can obtain fairly good estimates of the actual mean temperature profile on Mars. The basic reason for the insensitivity of the Martian surface temperature to changes in the carbon dioxide content and surface pressure is that, for the carbon dioxide amounts and surface pressures considered, the atmosphere is already opaque in the 15μ carbon dioxide band.

The thermal equilibrium model can also be applied at different latitudes and seasons to obtain a rough idea of the climatology of Martian surface and atmospheric temperatures. Figure 8 shows a pole to pole temperature cross-section for the Northern Hemisphere winter solstice on Mars that was computed by Ohring et al. (1967) with the thermal equilibrium model. The dashed curve represents the computed height of the tropopause, which marks the boundary between the convective lower layer atmosphere and radiative equilibrium upper atmosphere on Mars. The actual temperature cross-section for Mars would differ from this because of the effects of latitudinal transport of heat from the summer hemisphere to the winter hemisphere and, possibly, CO_2 condensation in the polar night region. However, the main features of the pole to pole cross-section — little or no latitudinal temperature gradient in the summer hemisphere and a greater latitudinal temperature gradient in the winter hemisphere — would still prevail, as indicated by the preliminary general circulation experiment for Mars that has been performed by Leovy and Mintz (1966). In comparison with the Earth's atmosphere, a major difference is the indication that the maximum summer temperature occurs at polar latitudes rather than at low latitudes, as in the case on Earth. Major reasons for this difference are the absence of a permanent ice cap on Mars and the greater transparency of the Martian atmosphere for solar radiation.

In a recent paper, Goody and Belton (1967) have computed radiative relaxation times for Mars. Radiative relaxation times are determined by the rate at which radiation can dissipate a thermal perturbation, which depends upon the fraction of the atmosphere that is composed of radiatively active molecules. On Earth this fraction varies from about 10^{-2} for water vapor at the surface in the tropics to 3×10^{-4} for carbon dioxide in the stratosphere. On Mars, because of its predominately carbon dioxide atmosphere, this fraction is greater than 0.5. Goody and Belton find that radiative relaxation times on Mars are shorter than on Earth by factors between 5 and 100 depending upon the scale of the perturbation. Thus, atmospheric radiative processes, which on Earth are important only for the longer time scales (>10 days) should be important on Mars even for shorter period phenomena. On the basis of their calculations, they conclude that: 1) diurnal changes of temperature will propagate to greater heights in the Martian

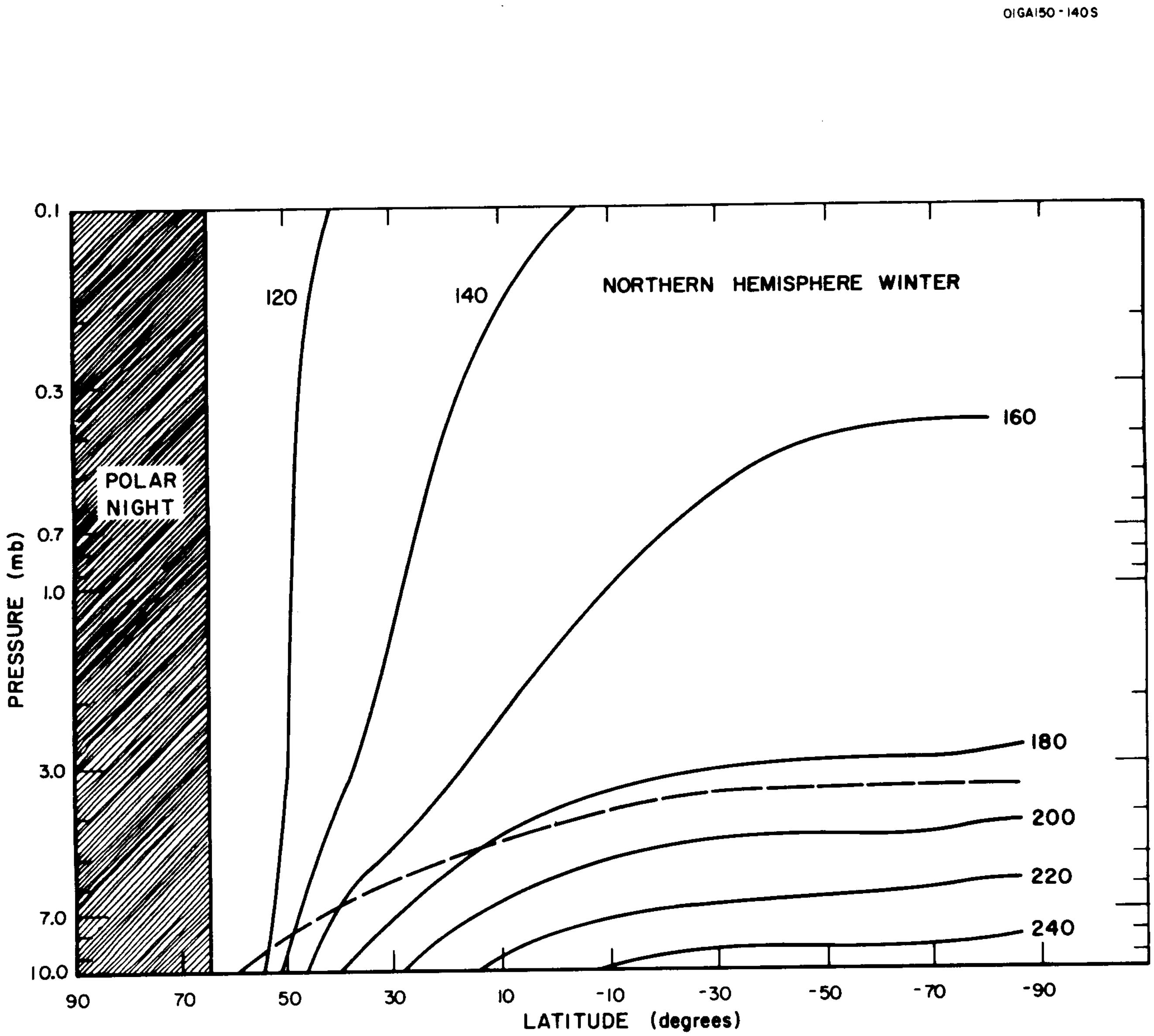

Figure 8. Latitudinal temperature cross-section for Mars based upon thermal equilibrium calculations (Temperature in ^{o}K). (After Ohring et al., 1967)

atmosphere than is the case on Earth, the propagation mechanism being mainly radiative transfer rather than turbulent diffusion; 2) A non-negligible diurnal wind system exists on Mars; and 3) non-linear interactions between the motion field and the initial state of radiative equilibrium are much smaller on Mars than on Earth. Thus, the influence of radiative processes on the dynamics of the Martian atmosphere appears to be much greater than is the case on Earth.

4. VENUS

The amount of information available on the Venusian atmosphere is extremely limited. Table 5 summarizes the available information on atmospheric parameters of importance to radiative calculations. The estimate of the surface pressure is uncertain by perhaps an order of magnitude. The atmospheric composition is unknown although carbon dioxide appears to be a major constituent and water vapor a minor constituent. The reason for the lack of information on the Venusian atmosphere is the presence of a uniform and perpetual cloud cover in the atmosphere, which hides the surface of the planet throughout much of the electromagnetic spectrum and makes interpretation of spectroscopic observations extemely difficult. This is especially so since the composition, height, thickness, and other characteristics of the cloud layer are still uncertain. Based upon its distance from the sun and an albedo of 0.73, the temperature that Venus would have in the absence of an atmosphere is 237°K.

TABLE 5. PHYSICAL PARAMETERS FOR VENUSIAN ATMOSPHERE

ALBEDO	SURFACE PRESSURE	COMPOSITION	CLOUDS	T_e
~0.73	~10 atm.	CO_2 + others	~100% cloud cover;	~237°K
		Small amount of H_2O above clouds	Composition, height, thickness, density, unknown	

Assuming that the clouds of Venus radiate as a blackbody in the infrared, Mintz (1961) calculated the radiative equilibrium temperature distribution above the clouds. Carbon dioxide, is an amount equal to 1000 m-atm above the clouds, is assumed to be the only important radiating gas, and the calculations are conducted with a simple model in which the infrared spectrum is divided into window regions and regions in which 100 m-atm of carbon dioxide absorb completely. An albedo of 80 percent is used in the calculations.

The resulting radiative equilibrium temperature profile above the clouds of Venus is shown in Figure 9. The computed temperature of the cloud top is 237°K. This is in good agreement with temperatures deduced from Earth-based observations of the 8 to 12μ infrared radiation from Venus. It is also in good agreement with T_e, the effective temperature of the net incoming solar radiation, which, for an albedo of 80 percent, is 220°K, suggesting a modest greenhouse effect for the atmosphere above the clouds. Other calculations of radiative equilibrium temperatures above the clouds of Venus have been carried out by Rasool (1963), who

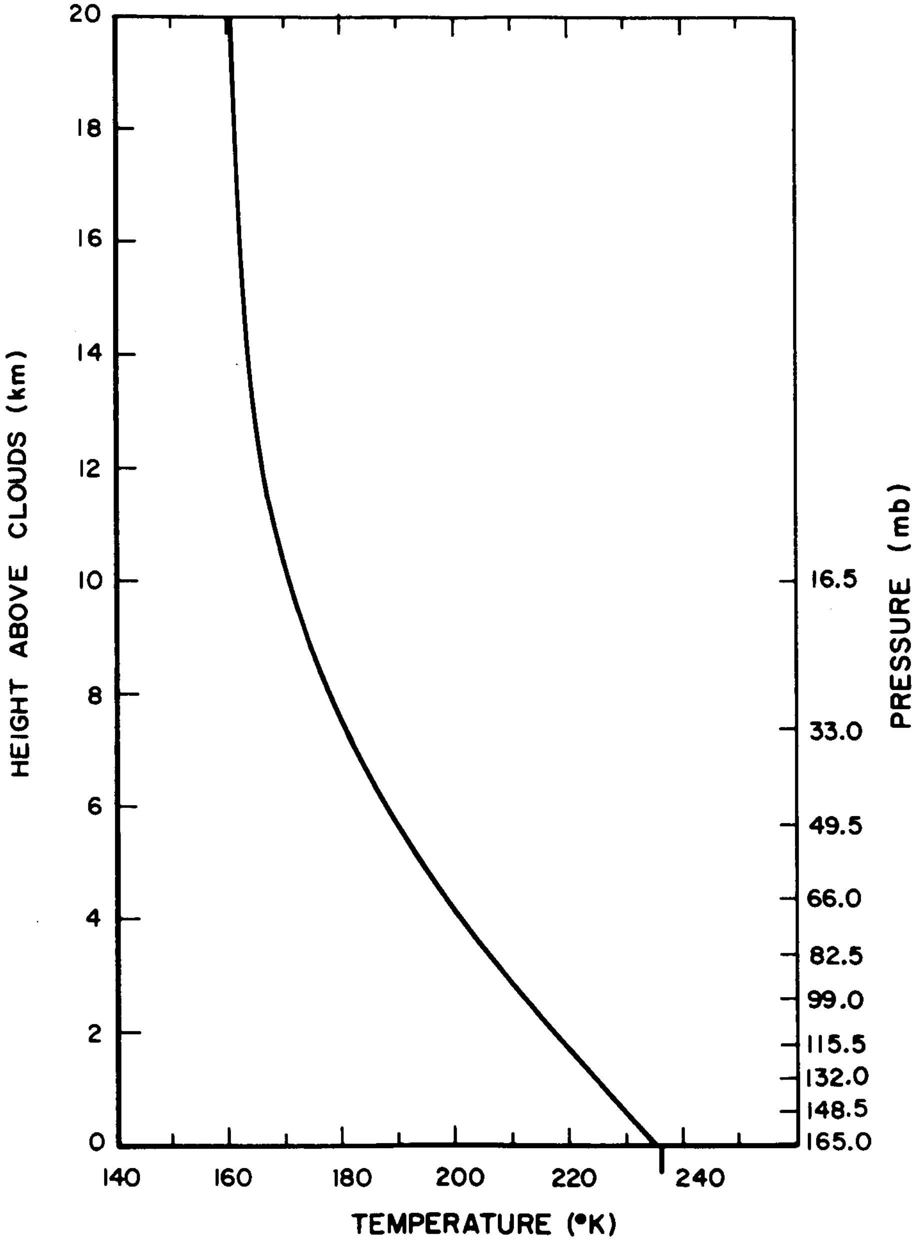

Figure 9. Radiative equilibrium temperature profile above the Venusian clouds (After Mintz, 1961).

assumed a grey absorbing atmosphere, and Hanel and Bartko (1964), who used a non-grey model.

Relatively little attention was given to the problem of estimating theoretically the surface temperature of Venus until one of the most surprising results of observational astronomy during the past decade became available. Earth-based radio telescope observations of the planet's emission of microwave radiation revealed that Venus radiates as a blackbody at about 600°K to 700°K at centimeter wavelengths. The Earth-based observations were confirmed by the microwave experiment on the Mariner II fly-by of Venus. Several sources have been proposed for the abnormal microwave emission, including ionospheric emissions, microdischarges between particles within the atmosphere, auroral radio noise, and the planet's surface. Newell (1967) has recently reviewed the proposed sources and attempted to analyze their validity. In this discussion, we shall be concerned with what most scientists believe to be the most probable explanation for the observed high microwave emission — a hot planetary surface. And, in particular, we shall discuss the following question: If the Venusian surface is really at a temperature of 600°K to 700°K, can this temperature be produced and maintained by a greenhouse effect.

A large greenhouse effect as an explanation for a high surface temperature on Venus was suggested by Sagan (1960). He used the equation shown in Table 6 for the balance of net incoming solar and outgoing infrared radiation at the top of the Venusian atmosphere. T_e is the effective temperature of the net incoming solar radiation, T_a is the effective radiating temperature of the atmosphere, T_s is the planetary surface temperature, and t is the transmissivity of the atmosphere for infrared radiation. The left side of the equation represents the net incoming solar radiation; the right side the outgoing infrared radiation. With this equation, Sagan computed the atmospheric infrared transmissivity required to maintain a 600°K surface temperature. Assuming T_a= 234°K, he obtained a required transmissivity of about 5×10^{-3}. Although low by terrestrial standards, the required transmissivity could be produced by amounts of carbon dioxide and water vapor that were not incompatible with information on the Venusian atmosphere. Sagan concluded that a large greenhouse effect was plausible. However, the transmissivity derived from Sagan's basic equation depends upon the choice of T_a, the effective radiating temperature of the atmosphere, and this temperature is really not known and must be assumed.

Another theoretical attack led to a different conclusion. Jastrow and Rasool (1963) used the Eddington approximation to compute the radiative equilibrium temperature profile of the Venusian atmosphere. For a grey atmosphere in radiative equilibrium, the Eddington approximation leads to the formula that is shown in Table 6 for the magnitude of the greenhouse effect. In this formula τ_s is the atmospheric opacity in the infrared. They found that, to maintain a surface temperature of 600°K, the atmospheric infrared transmittance must be less than 10^{-22}. A similar result with another grey atmosphere model was found by Ostriker (1963). Such low values were incompatible with estimates of amounts of absorbing gases on Venus.

Ohring and Mariano (1946) noted that the effect of an extensive cloud cover on the infrared radiation leaving the planet had not been specifically included

TABLE 6. GREENHOUSE MODELS OF THE VENUS ATMOSPHERE

INVESTIGATOR	BASIC EQUATION	REQUIRED GASEOUS TRANSMITTANCE
Sagan (1960)	$\sigma T_e^{\,4} = \left(\sigma T_a^{\,4}\right) + \left(t\sigma T_s^{\,4}\right)$	$\sim 5 \times 10^{-3}$
Jastrow and Rasool (1963)	$T_s/T_e = (1 + 0.75\tau_s)^{1/4}$	$< 10^{-22}$
Ohring and Mariano (1964	$T_s/T_e = \left\{(1 - n)\left[2E_3(\tau_s) + 2\tau_s^{-4k}\int_0^{\tau_s} \tau^{4k} E_2(\tau)d\tau\right] + n\left[2E_3(p\tau_s)p^{4k} + 2\tau_s^{-4k}\int_0^{p\tau_s} \tau^{4k}E_2(\tau)d\tau\right]\right\}^{-1/4}$	$\sim 10^{-2}$

in the two previous approaches. If the Venusian clouds have the property of being largely opaque to infrared radiation and relatively transparent to solar radiation, as are water clouds in the Earth's atmosphere, then perhaps the clouds plus the infrared absorbing molecules in the atmosphere could provide the required infrared opacity. To illustrate the effect of a cloud layer that is black in the infrared, Ohring and Mariano performed a series of computations with a simple radiative model. As the basic equilibrium condition, they assume a balance at the top of the atmosphere between net incoming solar and outgoing infrared radiation. The atmosphere is assumed to be a grey absorber, and the atmospheric temperature is assumed to decrease linearly with height. With these assumptions, an expression can be obtained that relates the magnitude of the greenhouse effect to the amount of cloudiness, n; the height of the cloud top; and the total infrared opacity of the atmospheric gases, τ_s. The resulting expression is shown in Table 6. In this equation, the E's are exponential integrals, p is the ratio of cloud-top pressure to surface pressure, τ is atmospheric infrared opacity, and k is proportional to the temperature lapse-rate.

With this expression, calculations of the magnitude of the greenhouse effect were performed for a variety of different cloud amounts, heights of the cloud top, and atmospheric infrared opacities. The magnitude of the greenhouse effect increases with increasing cloud amount, increasing cloud top height, and increasing atmospheric infrared opacity. For 99 percent cloud cover at a pressure level of a few-hundredths of the surface pressure, a gaseous transmittance of the atmosphere of the order 10^{-2} is required to maintain a 600°K surface temperature. This can be compared to the value of 10^{-22}, the required infrared transmittance of the atmosphere is the absence of a cloud layer.

Although these calculations indicate the possible importance of a cloud layer on the greenhouse effect, and make a large greenhouse effect seem more plausible, they do not prove that there is, indeed, a large greenhouse effect on Venus. For one thing, the model is somewhat artificial. For example, with a 100-percent cloud cover that is black in the infrared, the model would permit the surface temperature to be any value one desires, since the balance at the top of the atmosphere would be determined solely by conditions at the cloud-top and above. In addition, the existence of clouds that are fully black in the infrared is open to question. For cloud amounts approaching 100-percent coverage, small departures from blackness would influence the results obtained.

In work now in progress, Sagan and Pollack (1967) are carrying greenhouse effect calculations for Venus one step further. They are computing the greenhouse effect with a non-grey model atmosphere that contains a cloud composed of water substance. The absorption, transmission, and reflection properties of the cloud are determined from the assumed cloud particle densities and size distributions. They find that for a cloud top pressure of 0.1 atmospheres, a 600°K surface temperature can be maintained with an atmosphere consisting of 1-percent water, 10-percent carbon dioxide, and having a surface pressure of 10 atm. The details of the model and the results obtained are awaited with interest, for it is through calculations such as these that one can determine the validity of the greenhouse model as an explanation of the extremely high temperatures that apparently characterize the surface of Venus.

5. OUTSTANDING PROBLEMS

Despite the progress that has been made over the past several years on the applications of radiative transfer theory to the thermal structure of planetary atmospheres, a number of problems still remain. The treatment of clouds in radiative transfer problems is still highly simplified. Further research is required to develop techniques for computing radiative transfer within clouds on the basis of the cloud's characteristics- composition, particle size distribution, particle density, etc. The treatment of convection in thermal equilibrium models is highly empirical and does not deal with the basic dynamics of the problem. Further work should be directed toward developing a treatment of convection in terms of basic principles.

For the Earth's atmosphere, the problem of deriving the mean temperature profile for the troposphere and lower stratosphere appears to be under control. Further work is required on the problem of deriving the mean temperature profile of the upper stratosphere and mesosphere from considerations of radiative, photochemical, and convective equilibrium. Quantitative estimates of the changes in mean surface and atmospheric temperatures that would occur with changes in atmospheric composition or solar input can now be derived with some degree of reliability. Such estimates, and the theoretical models used for obtaining them, should be of great use to students of climatic change. The problem remains of determining the extent of changes in atmospheric composition or solar input that may have occurred in the past, or may occur in the future.

For Mars, the propagation of the diurnal temperature wave into the atmosphere and the interactions between radiative and dynamical processes require attention. For Venus, a major problem is the lack of such basic information as atmospheric composition, surface pressure, and cloud characteristics. This problem will be overcome as the planetary exploration program continues. When this information becomes available, it should be possible to make more reliable estimates of the greenhouse effect and of the thermal structure of the Venusian atmosphere.

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