Seasonal and Latitudinal Variations of the Average Surface Temperature and Vertical Temperature Profile on Mars

GEORGE OHRING AND JOSEPH MARIANO

GCA Corporation, Bedford, Mass.

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ABSTRACT

The seasonal and latitudinal variations of the average surface temperature and vertical profile of atmospheric temperature on Mars are computed using a thermal equilibrium model. It is assumed that carbon dioxide is the sole radiating gas in a model atmosphere that is composed of 60% carbon dioxide and has a surface pressure of 10 mb. The results are presented in the form of pole-to-pole temperature cross sections from the surface to about 40 km for each Martian season. The computed temperature cross sections indicate: 1) extremely small latitudinal temperature gradients in the summer hemisphere, with the maximum temperature occurring at the pole; 2) a decrease of tropopause altitude with latitude from a maximum at the equator during the equinoctial seasons and at the summer pole during the solstices; and 3) relatively isothermal vertical structure at high latitudes during the equinoxes and winter. Comparisons, where possible, of the present results with other theoretical studies and with the microwave observational indications of Martian temperatures yield generally good agreement.

1. Introduction

Previous theoretical estimates of the average vertical temperature profile in the lower Martian atmosphere have been based upon convective-radiative equilibrium models [see, for example, Goody (1957), Prabhakara and Hogan (1965), and Ohring and Mariano (1966)]. In such models, a surface temperature is assumed and the vertical temperature profile is computed on the basis of a convective troposphere and radiative equilibrium stratosphere. In the present study, we actually compute the average surface temperature as well as the average vertical temperature profile as a function of latitude and season. The computations are based upon the thermal equilibrium concept of Manabe and Strickler (1964), which is a slight modification of the convective-radiative equilibrium model. Although the effect on Martian temperatures of latitudinal transport of heat by the atmospheric circulation is not included in the model, the computed temperatures should be useful for many purposes. Where possible, we compare the computed temperatures with observational and previous theoretical indications of Martian temperatures.

2. Basic model

Under the concept of thermal equilibrium, the computed temperature profile must satisfy the following equilibrium conditions:

1) At the top of the atmosphere there is a balance between net incoming solar radiation and outgoing longwave 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 there is 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 superconvective 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.

The equations describing these equilibrium conditions may be found in Manabe and Strickler (1964). The thermal equilibrium temperature profile is computed with the use of an iterative procedure based upon the initial value method of computing radiative equilibrium temperatures.

Manabe and Strickler have successfully applied this concept to the earth's atmosphere. However, for the earth's atmosphere the critical lapse rate for convection can be taken as the observed mean tropospheric lapse rate of 6.5C km^{-1}. For Mars, we must assume a critical convective lapse rate, which, because of the dryness of the Martian atmosphere, we assume is the dry adiabatic lapse rate.

The radiative rates of temperature change depend upon two factors, infrared cooling and solar heating. Thus,

$$\left(\frac{\partial\theta}{\partial t}\right)_{\rm rad}=\left(\frac{\partial\theta}{\partial t}\right)_{ir}+\left(\frac{\partial\theta}{\partial t}\right)_{s},\tag{1}$$

where θ is temperature. For computation of the infrared cooling rates, the model of Rodgers and Walshaw (1966) is used, i.e.,

$$\left(\frac{\partial\theta}{\partial t}\right)_{ir}=-\frac{g}{c_p p}\left[B(Z)\frac{dT}{dz}(z,Z)-\int_0^Z\frac{dT}{dz}(z,z')\frac{dB}{dz'}(z')dz'\right],\tag{2}$$

where $(\partial\theta/\partial t)_{ir}$ is the infrared cooling rate at level z, where z, the vertical coordinate, is given by $z=-\ln\varphi$; $\varphi=p/p_0$, the ratio of atmospheric pressure at level z to the surface pressure; g is the gravitational acceleration; c_p the specific heat at constant pressure; B the blackbody flux for the absorption band; Z the highest level considered; and T the transmission function for the absorption band.

For the Martian atmosphere, we assume that only the 15μ CO_2 band contributes to the infrared cooling. Local thermodynamic equilibrium is also assumed. Curtis and Goody (1956) have found this to be a good assumption for the 15μ CO_2 band in the earth's atmosphere down to pressures as low as 3.4×10^{-2} mb. The Goody random model is adopted for the band transmission. Doppler broadening of the absorption lines is neglected on the basis of an analysis, following Rodgers and Walshaw (1966), which indicates that the transition from a Lorentz-line shape to a Doppler-line shape takes place at a pressure level of about 1.6×10^{-2} mb in the Martian atmosphere. For the CO_2 content and pressures prevailing on Mars, the expression for the band transmission can be written as

$$T=\exp\left[-\frac{(1.66\pi\bar{\alpha}k\bar{m})^{\frac{1}{2}}}{\delta}\right],\tag{3}$$

where $\bar{\alpha}$ is the mean half-width of an absorption line along an atmospheric path, k the mean line intensity, $\bar{m}$ the amount of CO_2 [gm cm^{-2}] along the path, δ the mean line spacing, and the factor 1.66 is introduced as a multiple of $\bar{m}$ to approximate flux transmission. The mean half-width between two levels is obtained with the use of the Curtis-Godson approximation, i.e.,

$$\bar{\alpha}=\frac{\alpha_s}{2}\left(\frac{\theta_s}{\theta}\right)^{\frac{1}{2}}\left(\frac{p_0}{p_s}\right)\left[\frac{\varphi(z)^2-\varphi(z')^2}{\varphi(z)-\varphi(z')}\right],\tag{4}$$

where α_s is the half-width at standard pressure p_s and standard temperature θ_s, and θ is a representative temperature of the Martian atmosphere, assumed to be 200K. The absorber amount between two levels includes a correction, following Rodgers and Walshaw (1966), for the effect of temperature on line intensities, i.e.,

$$\bar{m}=\frac{\Phi(\theta)wp_0}{g}[\varphi(z)-\varphi(z')],\tag{5}$$

where w is the mass mixing ratio of CO_2, and

$$\ln\Phi(\theta)=3.49\times10^{-3}(\theta-260)-1.28\times10^{-6}(\theta-260)^2,\tag{6}$$

where θ is 200K. The band parameters are $\alpha_s=0.07$ cm^{-1}, $k/\delta=718.7$ gm^{-1} cm^2, $\pi\alpha_s/\delta=0.448$, band center 667 cm^{-1}, and band width 170 cm^{-1} (Rodgers and Walshaw, 1966).

For the solar heating, we assume that only the near infrared bands of CO_2 contribute to heating. The method of Houghton (1963) is used for the computations. The average diurnal heating rate at a pressure level p is given by

$$\left(\frac{\partial\theta}{\partial t}\right)_s=\frac{g}{c_p}\overline{\cos\psi}\cdot r\sum_l I_{0_l}\frac{dA_l}{dp},\tag{7}$$

where $\overline{\cos\psi}$ is the average cosine of the solar zenith angle for the day, r the fraction of the day that the sun is shining, I_{0_l} the intensity of solar radiation per wavenumber at the top of the atmosphere in the lth absorption band, and A_l the integrated absorption of the lth band for the atmospheric column extending from the top of the atmosphere to the pressure level p along a slant path parallel to the solar beam, the summation extending over the CO_2 absorption bands. Houghton's (1963) expressions for the integrated absorption are:

$$\text{Weak bands}\quad(\bar{p}m<x_l):\quad A_l=a_l(\bar{p}m)^{\frac{1}{2}},\tag{8}$$

$$\text{Strong bands}\quad(\bar{p}m>x_l):\quad A_l=b+c_l\log_{10}(\bar{p}m),\tag{9}$$

where $\bar{p}=p/2$ [mb], m is the CO_2 path length [atm-cm], and a_l, b_l, c_l and x_l are constants for the lth band. The amount of solar radiation reaching the surface is given by

$$S_g=(1-A)r\,\overline{\cos\psi}(S_0-\sum_l I_{0_l}A_{gl}),\tag{10}$$

where A is the Martian planetary albedo, S_0 the intensity of solar radiation at Mars' distance from the sun, and the subscript g refers to the surface level. The empirical constants for the near infrared bands, after Houghton (1963), are shown in Table 1 along with the values of I_0 at Mars' mean distance from the sun.

3. Atmospheric composition, surface pressure and other input parameters

To perform the calculations with the thermal equilibrium model, information is required on the atmo-

TABLE 1. Constants for the near infrared CO_2 bands.

Band (μ)	I_0 [ergs cm^{-2} sec^{-1} $(cm^{-1})^{-1}$]	a	b	c	x
2.0	18.3	0.23	Weak only		
1.6	25.0	0.02	Weak only		
1.4	27.2	0.02	Weak only		
4.3	5.6	8.4	−11	37	15.4
2.7	11.8	1.87	−136	65	910

spheric composition and surface pressure on Mars. The results of recent spectroscopic observations (Spinrad *et al.*, 1966; Owen, 1966; Belton and Hunten, 1966) and the Mariner 4 occultation experiment (Kliore *et al.*, 1965) indicate that the surface pressure is about 5–10 mb and the atmosphere is predominantly CO_2. For our computations, we assume an atmosphere composed of 60% (by mass) CO_2 with a surface pressure of 10 mb. The remainder of the atmosphere is probably argon or nitrogen—or a combination of the two—both of which are inactive as absorbers of radiation. We assume that the remaining gas is all nitrogen.

Trace amounts of water vapor, $\sim 15\ \mu$ precipitable water, have been detected in the Martian atmosphere (Kaplan *et al.*, 1964). However, previous calculations (Ohring and Mariano, 1966) suggest that inclusion of this gas in our model would not substantially change the results. Therefore, it is not included in the computations.

Several other input constants are required in the computations. The intensity of solar radiation at the top of the Martian atmosphere is taken as $S_0 = \text{s.c.}/R^2$, where s.c.$=2$ cal cm^{-2} min^{-1} is the solar constant for earth and R is the Mars-sun distance in astronomical units. The variation of R with time of year is included in the calculations.

The planetary albedo of Mars is taken as 0.3 (de Vaucouleurs, 1964) and is assumed to be invariant with latitude and season. The convective lapse rate is assumed to be the dry adiabatic lapse rate. For the assumed atmospheric composition, the required thermodynamic constants are $(\gamma-1)/\gamma = 0.269$ and $c_p = 0.858 \times 10^7$ erg gm^{-1} $(°K)^{-1}$. The iterations continued until the net rates of temperature change were less than the prescribed convergence criterion of $\epsilon = 0.05$K day^{-1}.

4. Results and discussion

Computations were performed of the average surface temperature and atmospheric temperature profile on Mars as a function of latitude and season. A four-layer computational model was used after sample computations indicated no appreciable differences between a 10-layer and the 4-layer model. Temperatures were computed at even 20° latitude intervals. The computed temperature distributions are shown in Figs. 1, 2, 3 and 4 in the form of latitudinal cross sections, one for each season, extending from north pole to south pole.

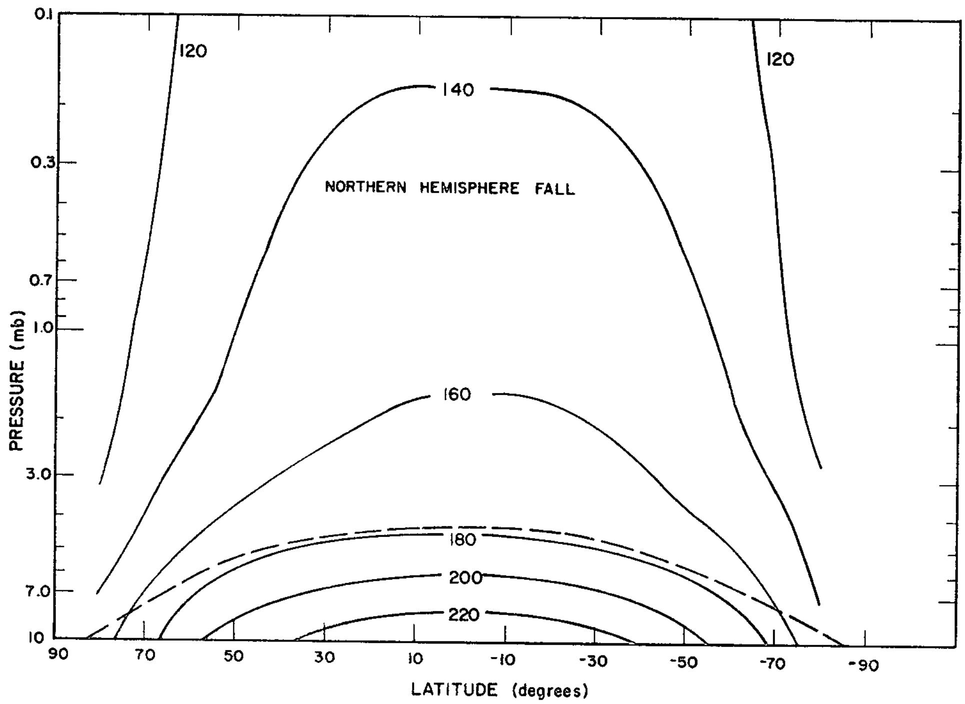

FIG. 1. Computed latitudinal temperature (°K) cross section for Northern Hemisphere fall equinox on Mars. Tropopause indicated by dashed line.

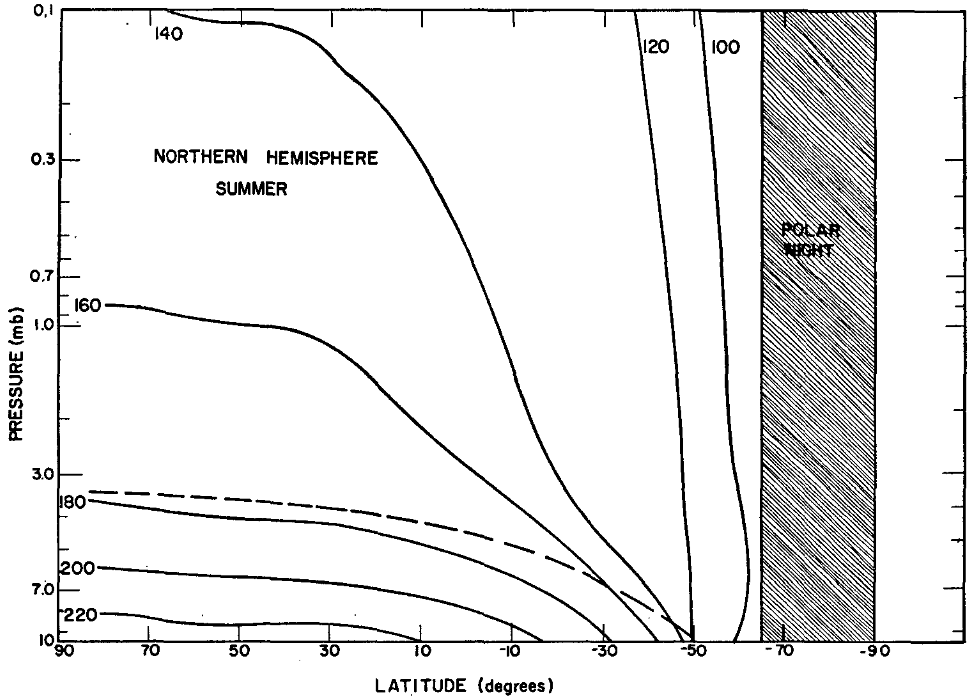

FIG. 2. Same as Fig. 1 except for summer solstice.

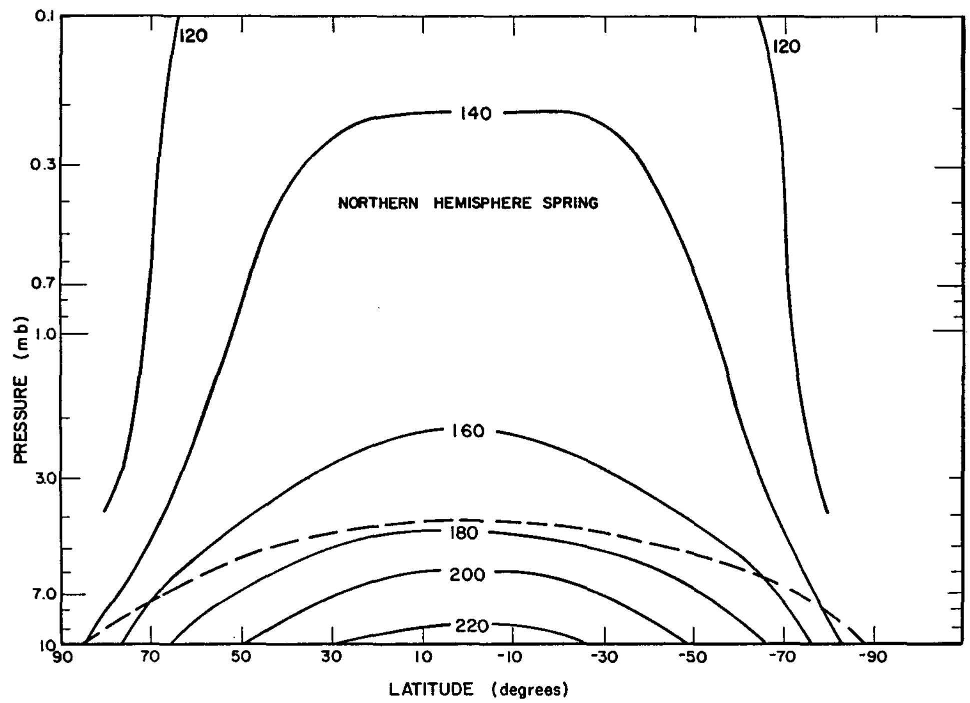

FIG. 3. Same as Fig. 1 except for spring equinox.

The vertical coordinate is the pressure plotted on a logarithmic scale; $p=1$ mb corresponds to about 20 km altitude, $p=0.1$ mb to about 40 km. The computed tropopause level is indicated by a dashed line.

The major features of the computed temperature cross-sections are: 1) the extremely small latitudinal temperature gradients in the summer hemisphere, with the maximum temperature occurring at the pole; 2) the decrease of tropopause altitude with latitude from a maximum at the equator during the equinoctial seasons

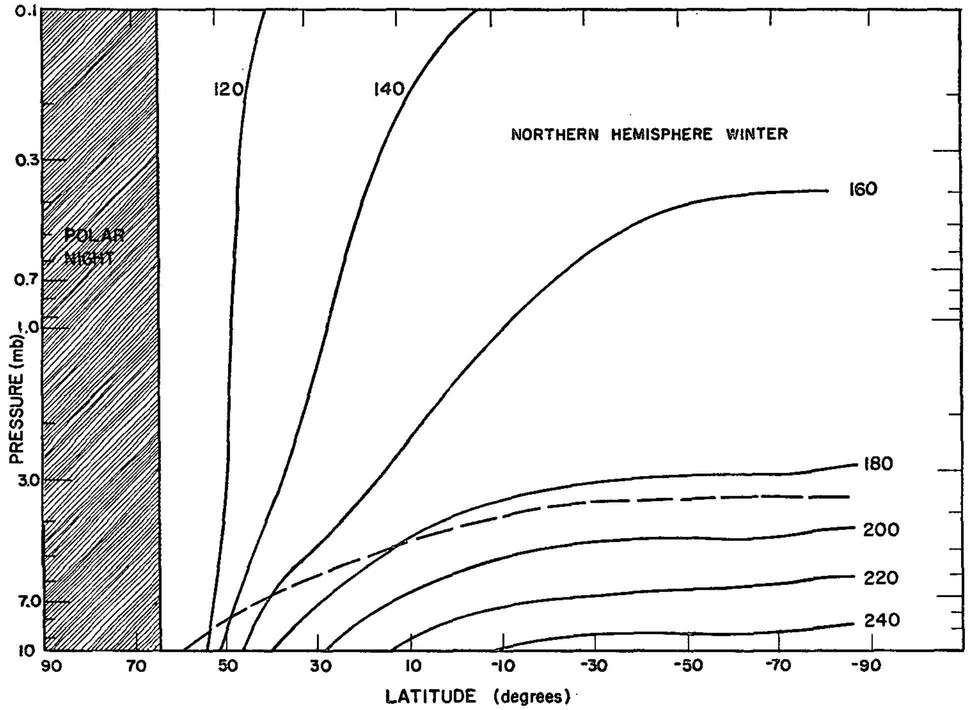

FIG. 4. Same as Fig. 1 except for winter solstice.

and at the summer pole during the solstices; and 3) the relatively isothermal vertical structure at high latitudes during the equinoxes and in winter.

The average planetary surface temperatures for the four seasons are listed in Table 2. The temperature differences between the two equinoxes and between the two solstices are due to differences in the planet's distance from the sun.

It is of interest to compare the computed temperatures with observational indications of Martian temperatures and with other theoretical estimates. Indications of surface temperature are available from microwave and infrared observations. Unfortunately, both the infrared and microwave observations are of the sunlit side of the planet and are therefore representative of daytime temperatures rather than the average daily temperatures that are computed in our model. However, the microwave observations refer to temperatures at levels a few centimeters below the surface. At such levels, the effect of diurnal temperature variations would be minimal, and the observed temperatures may be close to the daily average temperatures. Unfortunately, the microwave observations do not resolve the disk. Thus, to make a comparison we must average our computed surface temperatures to obtain a planetary average surface temperature. This temperature can then be compared with the microwave observations. Our computed planetary average surface temperature is 208K. The microwave observations yield an average temperature of about 215–220K, when corrected for a surface emissivity of 0.89 and normalized to a mean Mars-sun distance (Dent *et al.*, 1965; Hughes, 1966; Kellerman, 1966). The uncertainty in this temperature, due to observational errors, is about $\pm 5\%$. The agreement between the computed and observed planetary average temperature can be considered good.

Theoretical calculations of the latitudinal and seasonal distributions of the diurnal variation of Martian surface temperature have been performed by Leovy (1966) and Leighton and Murray (1966). Figs. 5 and 6 show comparisons of our computed surface temperatures with the daily average surface temperatures of Leovy. Leovy models the atmospheric radiative exchange in a fashion similar to ours. His treatment of convection is somewhat different, and includes the effect of conductive heat transfer between surface and subsurface. His computations are for a 5-mb surface pressure, 100% CO_2 atmosphere with an albedo that varies with latitude and whose average value is about 0.23. Because of a generally lower albedo, his surface temperature should be a few degrees higher than ours.

TABLE 2. Average planetary surface temperature.

Season	Average planetary surface temperature (°K)
Northern Hemisphere spring equinox	210
Northern Hemisphere summer solstice	194
Northern Hemisphere fall equinox	217
Northern Hemisphere winter solstice	212

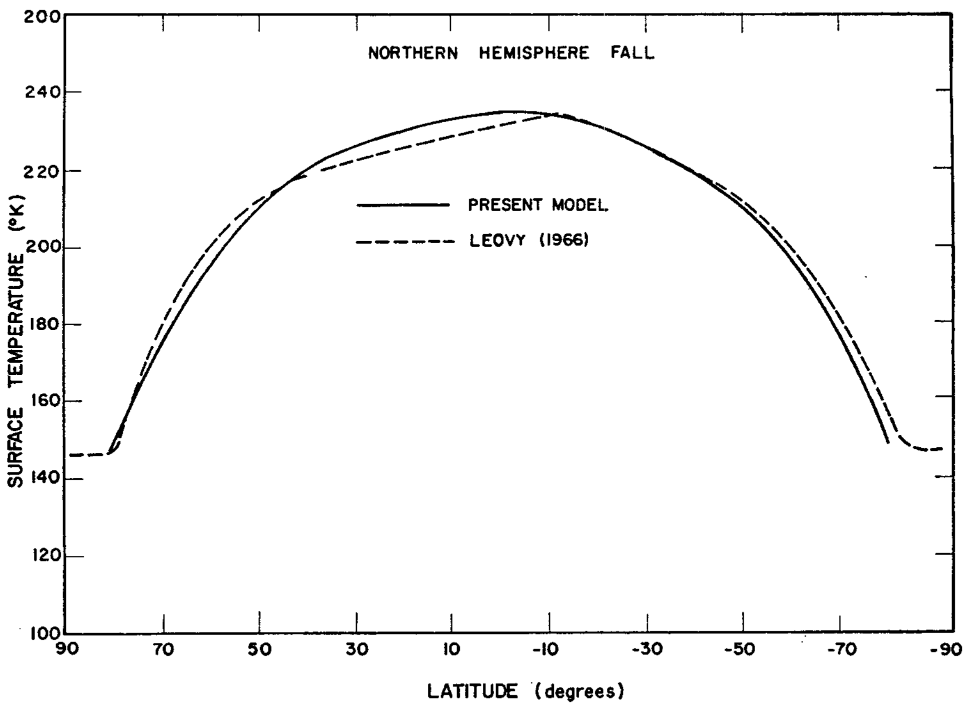

FIG. 5. Comparison of computed latitudinal variation of Martian surface temperature for Northern Hemisphere fall equinox with that of Leovy (1966). Solid line, present model; dashed line, Leovy (1966).

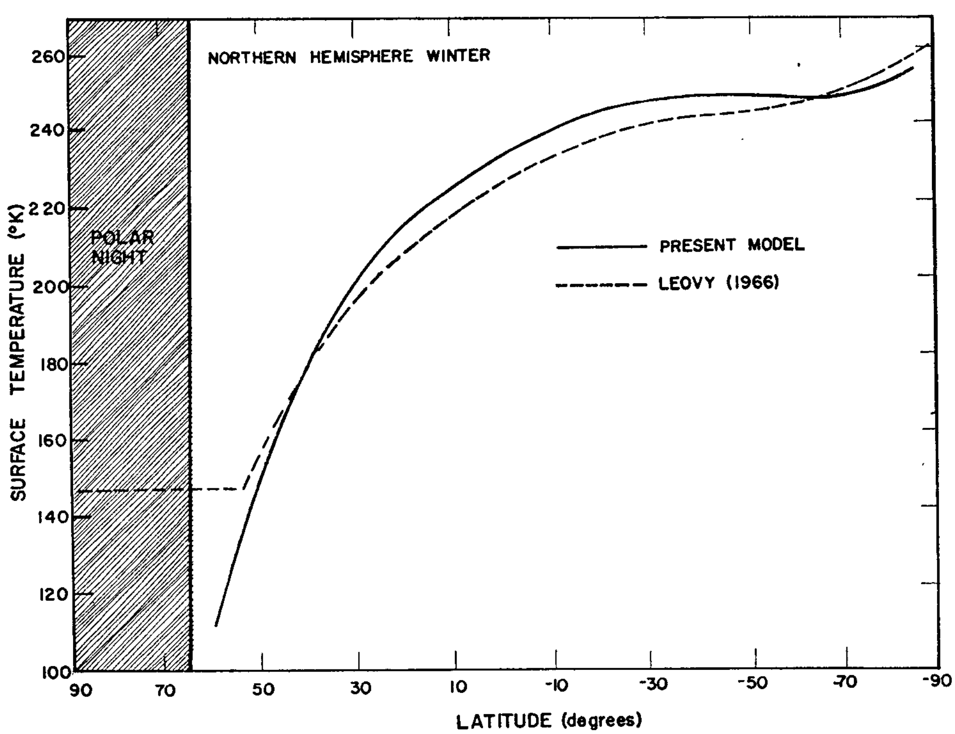

FIG. 6. Same as Fig. 5 except for winter solstice.

The results are in excellent agreement except at polar regions during the equinoxes and winter. When Leovy's surface temperature falls below the condensation temperature of CO_2 ($\sim$145K), he permits atmospheric CO_2 to condense. The release of latent heat of condensation maintains the polar temperatures during these seasons at 145K. This effect is not included in our computations.

A comparison of our computed latitudinal distribution of annual surface temperature with that of Leighton and Murray (1966) is shown in Fig. 7. Leighton and Murray parameterize the exchange of heat energy between the surface and atmosphere, but do include the effects of conductive heat exchange between surface and subsurface and the effect of the release of latent heat due to condensation of CO_2. Their assumed albedo is 0.15. The comparison indicates that their temperatures are some 8–10K lower than ours at low and middle latitudes. The results of all three theoretical models—ours, Leovy's and Leighton and Murray's—are in excellent agreement on the latitudinal variations of surface temperature.

There is no observational information available on the latitudinal and seasonal variations of the Martian vertical temperature distribution with which to compare our results. There are also no theoretical estimates except those from Leovy's (1966) two-layer atmosphere model. In Figs. 8 and 9 we compare our latitudinal distributions of computed temperature at the atmospheric level, $p/p_0=0.31$, where p is pressure and p_0 surface pressure, with those of Leovy for the same atmospheric level. In general, the computed temperatures are some 20–35K lower than Leovy's. This difference may be due to two factors. Leovy may have overestimated the mean diurnal convective heat flux, which would lead to higher atmospheric temperatures and lower surface temperatures in his model. For example, his computations are based upon the assumption that convective heat flux is predominantly upward with no downward convective flux permitted during the nighttime hours when turbulent exchange could be expected to transport heat from atmosphere to surface. This assumption leads to mean diurnal temperatures 8K higher at this level and about 15K higher at the level $p/p_0=0.77$ than they would be if both upward and downward heat flux were permitted (Leovy, 1966). This would also explain why surface temperatures agree quite well despite his lower albedos. Another factor may be the difference in 15μ band transmittance models. Leovy uses the transmission functions of Prabhakara and Hogan (1965). In a previous paper (Ohring and Mariano, 1966), we compared temperature profiles computed with a convective-radiative equilibrium model using Elsasser's (1960) transmission tables with temperature profiles computed by Prabhakara and Hogan (1965). Their temperatures were some 15–20K higher at upper levels. Our present temperature profiles, using the Rodgers and Walshaw (1966) transmission model, are in good agreement (<10K difference) with

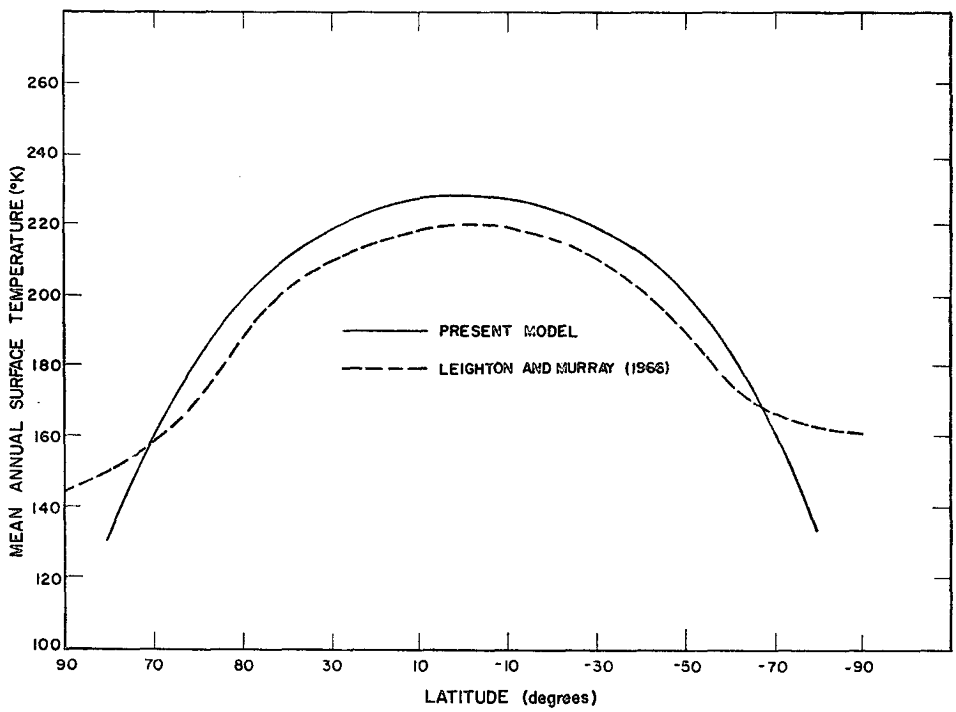

FIG. 7. Comparison of computed latitudinal variation of mean annual Martian surface temperature with that of Leighton and Murray (1966). Solid line, present model; dashed line, Leighton and Murray (1966).

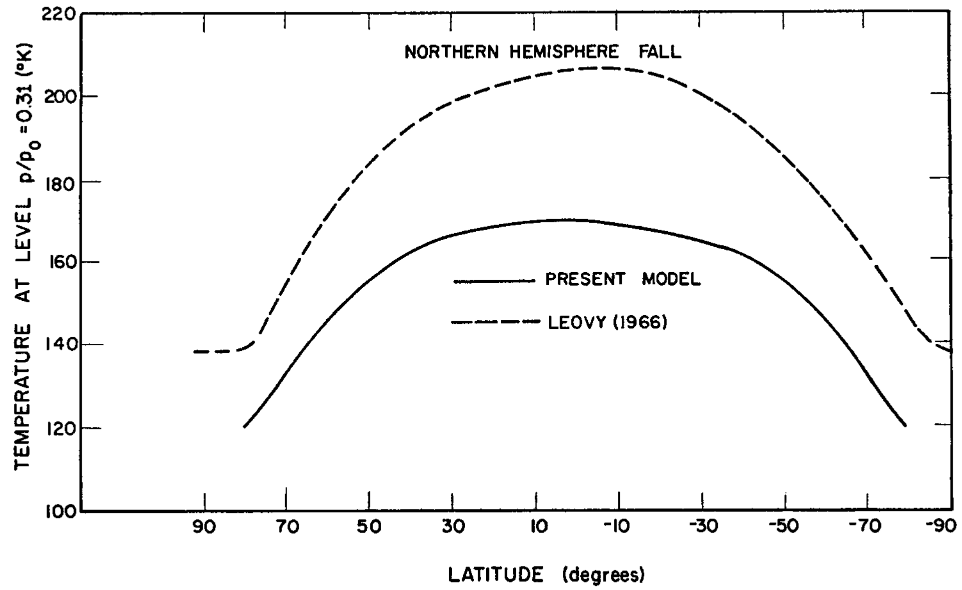

FIG. 8. Comparison of computed latitudinal variation of Martian temperature at atmospheric level $p/p_0=0.31$ for Northern Hemisphere fall equinox with that of Leovy (1966). Solid line, present model; dashed line, Leovy (1966).

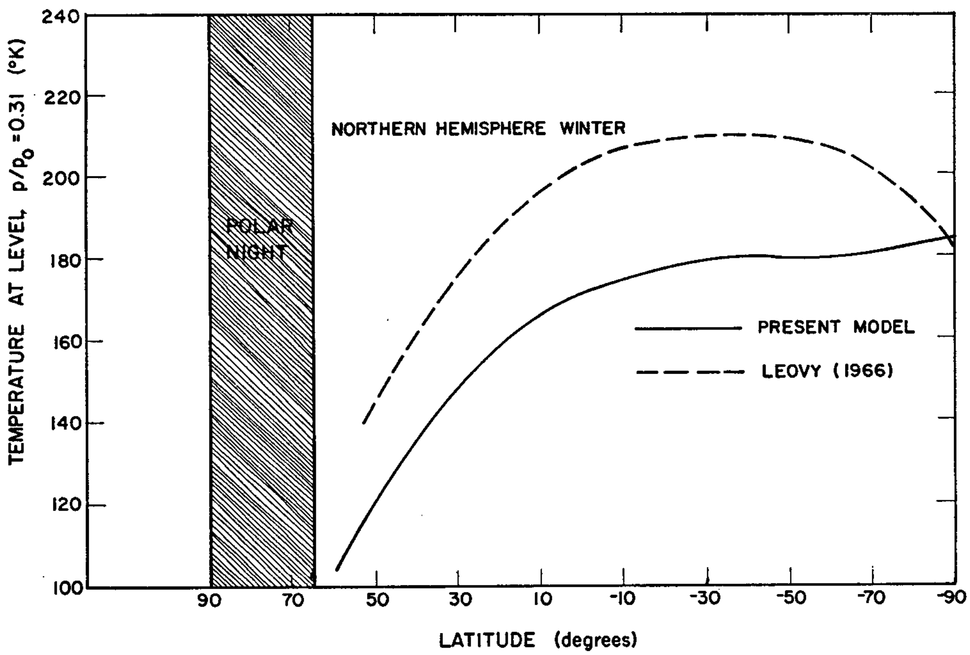

FIG. 9. Same as Fig. 8 except for winter solstice.

those computed with Elsasser's transmission data. This suggests that the use of different transmission models is another cause of the difference in results between the present paper and Leovy (1966).

5. Concluding remarks

Martian pole-to-pole temperature cross sections from the surface to about 40 km altitude for each season have been computed with a thermal equilibrium model. The computed average temperatures should be closest to the actual average temperatures at middle latitudes during the equinoxes and at low latitudes during the solstices. Because the model does not allow for latitudinal transport of heat energy by the atmospheric circulation and the possible condensation of carbon dioxide, the computed temperatures are too low at polar latitudes during the equinoxes and winter, and probably too high at equatorial latitudes during the equinoxes and at polar latitudes during summer.

Comparisons of the present results with other theoretical studies and with the microwave observations of Martian temperatures yield generally good agreement. Further theoretical studies of temperatures of the surface and lower atmosphere of Mars should include the effects of latitudinal heat transport, release of latent heat of condensation, conductive heat exchange between surface and subsurface, atmospheric water vapor radiative transfer, and analyses of the diurnal temperature wave.

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