Structure of Planetary Atmospheres

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Nomenclature

A = albedo, fraction of solar radiation reflected by the planet; total integrated albedo corresponds to the albedo over the whole solar spectrum (ultraviolet, visible, and infrared), whereas visual albedo applies only to the visible radiation
C = root mean square velocity of escaping particles
C_p = specific heat of a gas at constant pressure
G = universal gravitational constant
g = acceleration due to gravity
H = $k\bar{T}/\bar{m}g$, scale height
k = Boltzmann constant
M = mass of the planet
m = mass per particle
mb = 10^{-3} bar = 10^3 dyne/cm^2
N_e = electron density, electrons/cm^3
n_c = number density of the atmosphere which would exist at the surface if the atmosphere at all heights were isothermal at the temperature of the exosphere
n_0 = number density of the atmosphere at the surface, mole/-cm^3
R = radius of the planet
r = distance of the planet from the sun
$\bar{T}$ = mean temperature, °K
T_0 = temperature at reference level, °K
T_c = temperature at the level of escape or exosphere, °K
T_e = effective blackbody temperature of a planet $\propto r^{-1/2}(1 - A)^{1/4}$
T_G = temperature of the surface of planet
T_z = temperature at level z
t_e = time in which the atmospheric density of a planetary atmosphere will fall to $1/e \sim 1/2.7$ of its original value due to the gravitational escape of gases
z = altitude, km
β = adiabatic temperature gradient
$\varkappa$ = mass absorption coefficient, cm^2/g
ρ = density of the atmosphere at any given level, g/cm^3
ρ_0 = density of the atmosphere at reference level, g/cm^3
τ_0 = total optical thickness of the atmosphere = $\int_0^\infty \varkappa\rho dz$
τ_z = optical thickness of the atmosphere above the level z = $\int_z^\infty \varkappa\rho dz$

Introduction

INTENSIFIED research activity in the field of planetary atmospheres during the last few years has led to a number of new results that have forced a revision of our previous understanding of the structure of the atmospheres of Venus and Jupiter.

Recent measurements of the intensity of the radiation emitted by the planet Venus in the centimeter wavelength show that it corresponds to thermal radiation of temperature of $\sim$600°K.[1] Since radiation in the decimeter region probably passes unattenuated through the atmosphere and clouds of Venus, it generally is assumed that the measured temperature refers to the surface of the planet. Also, there is evidence that the atmosphere of Venus may not be composed predominantly of CO_2, as previously estimated,[2] but is present only as a minor constituent in a mixing ratio of 5%. The interpretation of these measurements has turned out to be one of the most interesting problems in planetary science at the present time.

It has been shown that Jupiter is the source of yet another type of intense radiation recently observed in the decimetric wavelength region.[1] There is also an indication that the Jovian atmosphere may be predominantly helium and not hydrogen, as so far has been believed.[3] In the case of Mars, renewed interest, both theoretical and observational, has led to revised atmospheric models.[4-6] The purpose of this article is to review the properties of the atmospheres of Mars, Venus, and Jupiter in the light of the most recent observational results.

General Considerations

Temperature, density, and composition are the three essential parameters that determine the structure of a planetary atmosphere. Of these, temperature is the most significant because it directly reflects the processes of energy absorption in the atmosphere and at the ground. The vertical temperature structure of the Earth's atmosphere is fairly well known,[7] and, therefore, before entering into a detailed discussion of the structure of the atmospheres of other planets, it perhaps will be desirable to understand the factors that determine Earth's observed temperature profile shown in Fig. 1.

Temperature

The solar radiation flux, with an effective blackbody temperature $\sim$6000°K, reaching the top of Earth's atmosphere, has a value of $\sim 1.4 \times 10^6$ erg/cm^2/sec. Part of this radiation immediately is "reflected" back to space by clouds and the atmosphere and does not play any role in the energy balance of the planet. The fraction of the solar radiation thus reflected back to space is known as the albedo of the planet, and, in the case of Earth, it has been estimated to be 0.39 (e.g., see Ref. 8). Most of the remaining 61% of the solar radiation,

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composed mainly of the visible part of the spectrum, penetrates down to the ground and heats the surface to a certain temperature denoted as T_e or the effective temperature of the planet. In the case of fast rotating planets (e.g., Earth, Mars, Jupiter),

$$T_e^4 \propto (SC/4)\cdot(1 - A)$$

where SC is the solar constant or the radiation flux received at the top of the atmosphere and A is the albedo of the planet over the whole solar spectrum. For the values of SC and A just given, the T_e for Earth is 245°K.

The surface radiating at this relatively low temperature emits primarily in the far infrared. A large fraction of the radiation emitted by the surface is absorbed immediately by molecules of CO_2 and H_2O present in the lower layers of Earth's atmosphere due to the excitation of intense vibration-rotation and pure rotation bands. A part of this absorbed energy in the infrared is returned to the surface and provides extensive heating of the ground, raising the surface temperature to the observed value of ~290°K.

This additional heating of the surface by the return of infrared from the atmosphere is referred to as the "greenhouse effect" and is an essential element in understanding the radiation budget of the planet in general and of the lower atmosphere in particular.

The lower atmosphere of Earth is in radiative and convective equilibrium, and the temperature decreases with height at a rate of ~ 6.5°K/km.

The adiabatic temperature gradient is given by $\beta_a = -g/C_p$, where C_p is the specific heat of air at constant pressure. For Earth's atmospheric composition, $\beta = -9.6$°K/km, but because of the eventual condensation of atmospheric water vapor in the form of clouds, the latent heat released in the atmosphere reduces the actual temperature gradient to the observed value of −6.5°K/km.

This region of negative temperature gradient extends up to an altitude of ~12 km and is referred to as the *troposphere*. The top of the troposphere is denoted as the *tropopause*, and the temperature at this level is of the order of ~200°K. Above the tropopause, however, the atmosphere is mainly in radiative equilibrium. Because the water vapor precipitates out at the low temperature of the tropopause, there is a very small amount of water vapor in the atmosphere above to provide any significant infrared opacity. The temperature therefore remains in the neighborhood of ~200°K and then rises subsequently because of the direct absorption of the solar ultraviolet radiation by the traces of ozone present at ~50 km.

The atmospheric region between the tropopause and the level of the secondary temperature maximum (at 50 km) therefore has a positive temperature gradient and is stable against convection. It is referred to as the *stratosphere*.

Above the ozone layer, the temperature decreases again with altitude to a minimum value of ~180°K at ~80 km (*mesopause*).

Above 100 km, a region of strong heating sets in which results from the photodissociation of O_2 and the photoionization of N_2 and O by the solar radiation in the far ultraviolet. This is the region of the *thermosphere*. At the top of the thermosphere, the temperature approaches a constant value of ~1500°K.

The *exosphere* is defined as the region where the atmospheric density and the probability of collisions is so small that the particles execute ballistic trajectories in the gravitational field of the planet. The base of the exosphere is a level above which there is only one collision for a particle moving vertically away from the planet. In the case of Earth, the base of the exosphere is located at an altitude of ~700 km, where the density is of the order of ~10^6 particles/cm². The exospheric region extends up to the distance where the atmospheric density falls to the value of ~100 particles/cm², the average density of the interplanetary medium. On this criterion, the outer boundary of the exosphere lies near 5000 km. This can be considered as the limit of Earth's neutral atmosphere.

Charged particles trapped in Earth's magnetic field, however, populate a region beyond this, called the *magnetosphere*, which has been found to extend to several Earth radii and is known more commonly as Van Allen radiation belt.

The structure of Earth's atmosphere, as outlined in the foregoing, may serve as a guide for understanding and describing the atmospheres of other planets. The different circumstances under which other atmospheres may have evolved, however, no doubt will lead to substantially different structures with new regions and layers whose relationships bare only slight resemblance to Earth's atmosphere. Nevertheless, if one can understand the basic physics of our own atmosphere, it is possible to go quite far in inferring the elemental properties of the atmospheres of other planets, taking into account differences in composition, planetary mass and radius, and distance from the sun, whose radiation spectrum plays a very important role in atmosphere structure.

Composition

The condensation of planets from a gaseous mixture of solar composition initially would have given a planet an atmosphere

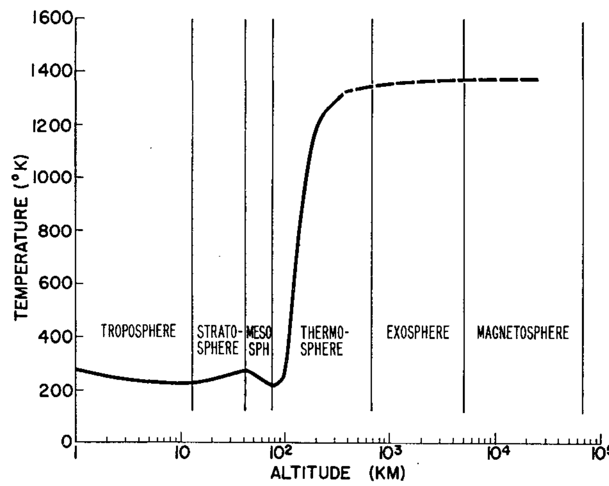

Fig. 1 Temperature profile of Earth's atmosphere (after Jastrow and Kyle, 1961[7])

composed predominantly of hydrogen and helium with relatively small amounts of CH_4, NH_3, H_2O, CO, O_2, N_2, CO_2, etc. Being the lightest of all elements, hydrogen and helium diffuse out to the fringes of the atmosphere and, depending on the size, mass, and temperature of the planet, eventually escape into the interplanetary medium. At any one time, therefore, the gross composition of a planetary atmosphere depends upon the steady state established as a result of the "escape velocity" of the planet and the mean thermal velocity of the atoms and molecules in the outer regions of the atmosphere. The atmospheric constituents are replenished by the exhalation of gases from the crust, and to some extent its composition is also modified by the solar ultraviolet radiation through the photodissociative and ionizing processes usually prevalent in the upper atmosphere. Factors like the chemical reactions of the gases in the atmosphere with the solid material of the crust, the capture of gaseous constituents from the interplanetary medium, and the radioactive processes in the interior of the planet also may contribute to the composition of the atmosphere.

From the considerations of Jean's theory of gravitational escape of gases, one can grossly estimate the present composition structure of Earth's atmosphere. According to Spitzer,[9] the time t_e in which the density of an atmospheric constituent

Table 1 Sea level composition of Earth's atmosphere

Constituent	% volume	Molecular wt (O = 16.00)
N_2	78.09	28.016
O_2	20.95	32.000
A	0.93	39.944
$(H_2O)^a$	0 to 2.0	18.016
CO_2	0.03	44.010
Ne	1.8×10^{-3}	20.183
He	5.24×10^{-4}	4.003
Kn	1.0×10^{-4}	83.7
H_2	5×10^{-5}	2.016
N_2O	2.5×10^{-5}	44.032
CO	2.0×10^{-5}	28.010
Xe	8.0×10^{-6}	131.3
CH_4	5×10^{-6}	16.035
$(O_3)^b$	0 to 7×10^{-6}	48.000

[a] Above the troposphere, the amount of H_2O is always very small.
[b] O_3 reaches a maximum of 1 to 3×10^{-4}% between 20 and 30 km.

of molecular or atomic weight m will fall to e^{-1} of its original value is given by

$$t_e = [B(6\pi)^{1/2}C/3g](e^y/y) \tag{1}$$

where $B = n_0T_0/n_cT_c$ and $y = GmM/kT_cR$; n_0 and T_0 are the particle number density and temperature, respectively, at the ground level, whereas n_c and T_c are the same parameters at the escape level, which, as described before, is the base of the exosphere; R is the radius of the planet, and C is the root mean square velocity of the escaping particles.

For an exospheric temperature of ~1500°K, hydrogen would have escaped from Earth in ~10^5 yrs. As the age of the planets is ~3×10^9 yr, the absence of hydrogen in the atmosphere of Earth is comprehensible. Hydrogenated gases like CH_4 and NH_3, which are susceptible to dissociation by the solar ultraviolet radiation, also would lose their hydrogen, and the carbon and nitrogen thus liberated probably will remain in the atmosphere as CO_2 and N_2.[10]

The case of water is different. Because of the low temperature of the tropopause, almost all of the terrestrial water vapor remains confined to the troposphere. An extremely minute amount, corresponding to the saturated vapor pressure of 200°K, will traverse the tropopause and will be susceptible to dissociation in the upper atmosphere. This special circumstance, according to Urey,[10] preserves the water on our planet. Urey also has argued that the present amount of free oxygen can be accounted for by the amount of water vapor so far dissociated in the upper atmosphere.

In the case of He, the time of escape from Earth is of the order of 10^8 yr, which, because of the possible uncertainty in the assumed escape level temperature, cannot be interpreted as a definite indication of complete absence of He in Earth's atmosphere. In fact, a layer of helium at an altitude of ~1600 km recently has been detected by satellite investigations.[11]

From Eq. (1), it can be concluded that exospheric constituents of molecular or atomic weight >6 probably will be retained by Earth up to the present time. These considerations indicate that H_2 has escaped, and hence the present atmosphere of Earth should be oxidized state. Table 1 shows the observed composition of Earth's atmosphere.

Nitrogen, oxygen, and argon make up more than 99.9% of Earth's atmosphere. Though the almost complete absence of hydrogen and helium perhaps can be explained by the gravitational escape of gases, the presence of free oxygen and the relative absence of CO_2 in the terrestrial atmosphere are problems related to the presence of life on Earth and the possible reactions of atmospheric gases with the crust. This subject very recently has been discussed comprehensively by Urey[10] and therefore will not be treated here in any detail.

The mean molecular weight for the atmospheric composition given in Table 1 is ~29. Earth's atmosphere is supposed to be mixed up to an altitude of ~80 km, and the composition remains nearly constant except for H_2O and O_3, as indicated in the table.

The total pressure at the surface due to this atmosphere is ~10^6 dyne/cm² (1000 *mb*) and the number density at the ground ~2×10^{19} mole/cm³. In a nearly isothermal atmosphere, the variation of density with height is given by the formula expressing hydrostatic balance:

$$\rho = \rho_0 e^{-[(h - h_0)/k\bar{T}]\bar{m}g} \tag{2}$$

in which ρ and ρ_0 are the densities at height h and h_0, $\bar{m}$ is the average molecular weight per particle, $\bar{T}$ the average temperature between h and h_0, g the acceleration of gravity, and k the Boltzmann constant.

The quantity $k\bar{T}/\bar{m}g$ is known as the scale height H of the atmosphere; at an altitude of one scale height, the density being reduced by a factor of e. For a mean temperature of 250°K in the first 80 km of the atmosphere, the atmospheric scale height is ~8 km.

Above 120 km, diffusive separation sets in and the pressure of each constituent varies in accordance with a scale height calculated for its own molecular weight. At the boundary of this domain, at a height of 120 km, nitrogen and oxygen are the major constituents, with N_2 dominating in the ratio 4:1, as on the surface. At higher altitudes the relative concentration of oxygen rapidly increases, and by 300 km the surface proportions are reversed, with oxygen now dominating in the ratio of 3:1.

The reason for this circumstance is that O_2 can be dissociated by absorption of solar ultraviolet, with relatively high probability, yielding atomic oxygen with half the weight per particle and, therefore, with twice the scale height and a correspondingly smaller rate of decrease of concentration with increasing altitude. Molecular nitrogen also undergoes photodissociation, but with a much smaller probability than for oxygen; hence it remains in primarily molecular form up to greater altitudes. Thus, again the twice-as-heavy nitrogen settles out of the atmosphere with relative rapidity. At 300 km, for example, less than 1% of the N_2 is in atomic form, whereas 99.5% of the oxygen is atomic.

For this reason, atomic oxygen becomes the principal atmospheric constituent above 200 km and continues to dominate the composition of the upper atmosphere up to about 1000 km. Above that level, the lighter gases, hydrogen and helium, emerge as the principal constituents because of their very large scale heights, although they are present only in trace amounts at lower altitudes. At the greatest altitudes, all gases except hydrogen have settled out of the atmosphere, and this lightest gas dominates, until finally the hydrogen atmosphere merges into the interplanetary medium at a distance of some 2 to 10 Earth radii. Figure 2 shows the variation of atmospheric composition with altitude.

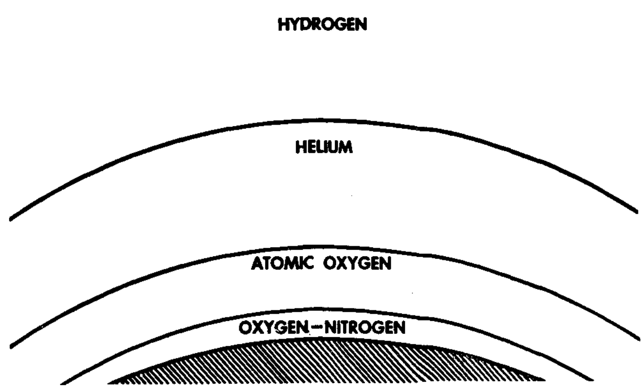

Fig. 2 Composition of the upper atmosphere of Earth (after Jastrow, 1962)

Other Planets

It is interesting to use one's knowledge of the physical properties of Earth's atmosphere to interpret the observational results obtained for the other planets. In a detailed analysis of the problem, Urey[10] has surveyed the properties of the atmospheres of the other planets in the light of observational evidence available until 1958 and has arrived at many interesting conclusions regarding the origin of planetary atmospheres. Using improved techniques of optical and radio astronomy, many new observational results have been obtained in the past four years. In order to fit the observed data, new model atmospheres for Venus, Mars, and Jupiter have been derived. The properties of the atmospheres of these planets therefore will be discussed in more detail.

Mars

Composition and Pressure

Table 2 summarized reliable physical data for Mars. From the considerations of the cosmic abundance of elements, the gravitational escape of gases, and the temperature environments of Mars, Kuiper[12] has given a list of gases which probably could be present in the atmosphere of Mars; they are given as follows in order of predicted relative abundance: $CO_2 + CO$; (NO); $N_2(N_2O)$; (COS); (SO_2); H_2O; $O_2 + O_3$.

The gases shown within parentheses may have been dissociated photochemically and hence their abundances reduced greatly. The rare gases like argon, xenon, and neon also would be present, but their abundances are very uncertain, especially that of argon, because it is generated continuously by radioactive decay of K_{40}. Keeping in mind these possible atmospheric constituents, the observational evidence of the composition of Martian atmosphere will now be discussed.

The only gas that has been detected spectroscopically up to now in the atmosphere of Mars is carbon dioxide.[12] Its abundance has been estimated by Goody and Grandjean[13] as 35 m-atm (where m-atm is the thickness of a homogeneous atmosphere in meters at 0°C and 760 mm). Thorough search for water vapor has failed to give a positive result, and recent spectroscopic observations by Kiess et al.[14] set an upper limit of 8×10^{-3} g/cm^2. From estimates of the thickness and dimensions of the polar caps, de Vaucouleurs[15] estimates the amount of water vapor at any time in the atmosphere of Mars to be $\sim 10^{-3}$ g/cm^2, which is below the amount observable from Earth, even from high altitude balloons. Kuiper[12] has looked for SO_2, O_3, N_2O, CH_4, and NH_3 but has failed to detect them, giving the following upper limits to the possible abundances of these gases in the Martian atmosphere:

N_2O	2.0 m-atm
NH_3	0.2 m-atm
CH_4	0.1 m-atm
O_3	5×10^{-4} m-atm
SO_2	3×10^{-5} m-atm

Dunham[16] estimates the amount of O_2 on Mars to be less than 0.15% of the amount on Earth, which gives an upper limit of 2.4 m-atm. Recently Sinton[17] has studied the absorption spectra of Mars in the 3 μ region and has given an upper limit to the amount of NO_2 and N_2O_4 as 2.2 m-atm in the atmosphere of Mars.

Table 2 Astronomical data for Mars

Mean distance from sun	1.52 a.u.
Mean equatorial diameter	6790 km
Length of day	1.0012 Earth day
Length of year	1.8808 Earth yr
Mass	0.1078 (Earth 1)
Mean density	3.90 g/cm^3
Gravity	377 cm/sec^2
Total integrated albedo	0.26 ± 0.02 (Ref. 81)
Effective blackbody temperature	209°K

Table 3 Probable composition of the Martian atmosphere

Gas	Amount, m-atm	% volume
N_2	$\sim$1675	95.0
A	$\sim$50	2.5
CO_2	$\sim$35	2.0
O	< 2.4	< 0.15
H_2O	$< 2 \times 10^{-3}$ g/cm^2	...

De Vaucouleurs[15] observed the brightness of Martian features at various angles and, from the amount of light scattered, estimated the surface pressure of Mars to be 80 ± 13 *mb*. Dollfus,[18] by polarization measurements, finds the surface pressure to be approximately 85 *mb*. The most likely value, according to Urey,[10] is 85 ± 10 *mb* which amounts to an atmospheric mass per unit area of 230 g/cm^2, corresponding to a total atmosphere of 1760 m-atm.

If one assumes 35 m-atm of CO_2, then the remaining unidentified constituents of the atmosphere are probably mostly nitrogen with traces of argon. These gases lack absorption spectra in the observable part of the solar spectrum but are abundant cosmically.

The probable composition of the Martian atmosphere is shown in Table 3. This atmosphere has a mean molecular weight of 28.5. With a surface pressure of 85 *mb* and an approximate temperature of 210°K, the surface density would be approximately 2×10^{18} particles/cm^3, which, for an isothermal atmosphere, should decrease exponentially with altitude with a scale height of $\sim$20 km. The vertical distribution of pressure for this model has been calculated by Goody[19] and is shown in Fig. 3. Results of a similar computation for Earth, assuming an isothermal atmosphere at 250°K, also are shown in Fig. 3 and are in accord with recent rocket measurements. It is interesting to note that, although the pressure at the surface of Mars is about $\frac{1}{12}$ of that of Earth, at an altitude of 31 km the two atmospheres have the same pressure, and above this height the pressure in the Martian atmosphere is greater than the pressure at the corresponding height in Earth's atmosphere. Because the decrease of density with height in the atmosphere of Mars is almost $2\frac{1}{2}$ times slower than in Earth, the levels of the ionosphere and thermosphere on Mars would be much higher than on Earth. In Fig. 3, Goody has indicated the corresponding levels at which photochemical reactions take place.

Temperature

Planetary temperatures usually are estimated by measuring the infrared radiation emitted by the planet. A large part of the infrared spectrum, however, is absorbed by the water vapor and CO_2 present in Earth's atmosphere. Ground-based observations of the planets in the infrared are therefore confined in the 8 to 12 μ region where Earth's atmosphere is relatively transparent. This spectral region in the infrared is therefore known as the atmospheric "window."

Extensive temperature measurements of Mars by infrared radiometry in this "window" of Earth's atmosphere have been made since 1926, and the results to date are quite consistent. Since the Martian atmosphere also should be largely transparent in the 8 to 12 μ "window," except for the weak absorption by CO_2 at 9.4 and 10.4 μ, it is assumed that the radiation intensity measurements in this wavelength region refer to the surface of the planet. Adel[20] and recently Hess,[21] however, have pointed out that the presence of relatively large amounts of CO_2 in the Martian atmosphere makes all measurements of temperature on Mars doubtful, especially those near the limb of the planet. The argument is that, as the atmosphere probably will be cooler than the surface,

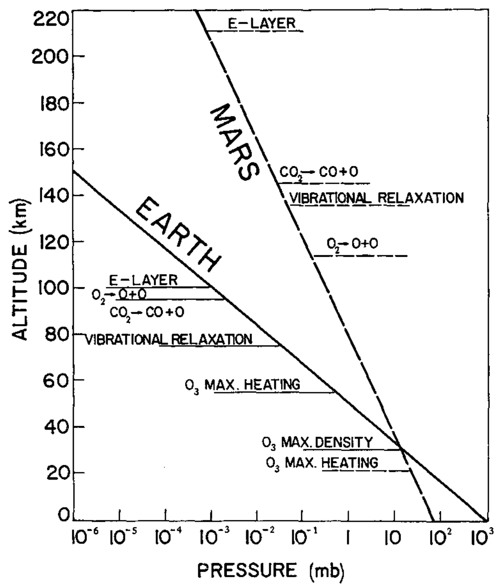

Fig. 3 Pressure vs altitude for Mars and Earth (after Goody, 1957[19])

emission from the CO_2 bands in the "window" will make the measured temperatures lower than the surface temperature.

From the known intensities of CO_2 bands at 9.4 and 10.4 μ,[22] one can estimate the contribution of CO_2 in the absence of water and other absorbing gases. Assuming a surface temperature of 300°K, the correction due to 35 m-atm of CO_2 at an average pressure of ~50 *mb* and at a mean temperature of 240°K is less than 3°K. One therefore can assume that, except for measurements made at limb, the temperatures measured in the 8 to 12 μ window refer to the surface of the planet.

De Vaucouleurs[23] has summarized all the available temperature measurements made until 1952, and they appear to be consistent within the possible observational error. Sinton and Strong[24] recently have repeated these observations obtaining similar values. Sinton and Strong and some of the earlier investigators also were able to scan the planetary disk latitudinally and longitudinally so as to obtain the diurnal and seasonal variation of temperature as a function of latitude. The up-to-date information can be summarized as follows:

Maximum temperature at equator	~300°K
Mean amplitude of diurnal variation	
Noon to sunset	~60°K
On earth in desert	~30°K
Night side temperature cannot be measured but probably can be estimated at the equator	~200°K
Day side temperature at poles	~220°K
Mean temperature of day side	~260°K
Mean temperature of whole planet	~230°K

The mean temperature of the day side of the planet should be compared with the temperature values obtained by radio measurements at 3-cm wavelength. Mayer, McCullough and Sloanaker[25] found an apparent blackbody disk temperature for Mars of 218° ± 50°K. Later Giordmaine et al. using Maser techniques,[26] made a more precise measurement of 211° ± 20°K. According to Mayer,[1] the radiation at radio frequencies probably is emitted from a few centimeters beneath the surface of the planet. As these measurements refer to the day side of the planet, the region just below the surface will be cooler than the surface itself, which will explain the discrepancy between the infrared and radio measurement.

Sinton and Strong[24] also have confirmed the earlier results regarding the difference in temperatures between the dark and bright regions on Mars, the dark region being warmer by ~8°K.

Vertical Distribution of Temperature

The atmosphere of Mars is optically thin in the infrared, and the probable absorbing gases are CO_2 and H_2O, which also are present in Earth's atmosphere. The vertical temperature profile in the atmosphere of Mars, therefore, cannot be determined by observation from the surface of Earth. The measurement is feasible from space vehicles by observing at the center of the strong CO_2 bands, and such experiments may be expected in the future. Only theoretical estimates, therefore, exist regarding the vertical temperature structure of Mars, but the models thus far derived vary considerably from author to author. Urey[10] has summarized and commented upon the earlier results of Hess[27] and Goody[19] and has given his own estimate of the height of the tropopause on Mars, ~30 km.

For the composition of Martian atmosphere, as given in Table 2, the adiabatic temperature gradient will be ~−3.7°K/km. For an equatorial ground temperature of 300°K, Urey therefore obtains a tropopause temperature of 187°K. Goody[19] had previously treated two model atmospheres for Mars, the first in which only CO_2 is the infrared absorber, and the second where 10^{-2} g/cm^2 of water vapor is responsible for the infrared opacity. The ground temperature was assumed to be 270°K. The tropopause in the two cases was estimated to be at 8.5 and 25 km, respectively. It is not known which of the two models is closer to the real Martian atmosphere. In fact, Jastrow and Rasool[28] recently have shown that, in order to explain the observed mean surface temperature of 230°K by a greenhouse effect, the infrared absorption by the atmosphere must be greater than what can be accounted for even by the inclusion of the specified amounts of both CO_2 and water vapor. If both of these gases are present, the Martian tropopause may lie at an intermediate level between 9 and 25 km.

Arking[4] recently has calculated a model atmosphere for Mars, allowing for convection and using the exact equation of radiative transfer for frequency independent absorption. A total optical thickness of 0.5 was chosen to obtain a surface temperature of 235°K, consistent with the observed mean surface temperature. Assuming an effective blackbody temperature of 217°K, an adiabatic gradient of −3.7°K/km in the convection zone, and an exponential dependence of absorption on altitude with a scale height of 17 km, the temperature profile shown in Fig. 4 (insert) is obtained. The convection zone is found to extend up to 8 km.

Ohring[6] also has investigated recently the vertical temperature profile for a model Martian atmosphere containing 2% CO_2, 98% N_2, and no water vapor. The ground temperature was assumed to be 230°K, and the transfer of radiation in the atmosphere was calculated for frequency-dependent absorption by the CO_2. The tropopause in this case was found to be at 9 km at a temperature of 196°K. In the stratosphere, the temperature keeps on decreasing and reaches a value of as low as 90°K at an altitude of 42 km, where the total pressure is 2.5 *mb*. As pointed out by Ohring himself, at such low stratospheric temperatures the atmospheric CO_2 will freeze out. The author attributes CO_2 cloud layer thus formed to the observed phenomenon of "blue haze" in the Martian atmosphere.

On the basis of arguments developed in a later section, this explanation seems unacceptable, and the stratospheric temperatures in Mars are probably higher than the frost points of CO_2 at the corresponding pressure levels. At 42 km altitude, carbon dioxide will condense at a temperature of 140°K.

The forementioned calculations of the temperature distributions in the Martian atmosphere do not take into account possible heating of the lower atmosphere by direct absorption of solar radiation in the ultraviolet. In the case of the terrestrial atmosphere, the ozone heating produces a temperature maximum at 50 km, but in Mars, with much less oxygen as compared to Earth, ozone would be confined to lower layers of the atmosphere.[28–30] The solar radiation in the region of 2500 Å responsible for ozone dissociation and atmospheric heating will penetrate to a much lower depth in the Martian atmosphere. Because of a higher atmospheric density at this level (compared to the density in Earth's atmosphere at 50 km) and because of the lower intensity of the solar radiation at the distance of Mars, the heating rates probably will not be as important as in the ozonosphere on Earth. However, they may affect the temperature gradients substantially and limit the extent of the convection zone.

Surface Features of Mars

As seen by telescope, the outstanding features of Mars are 1) the dark areas (maria), 2) the general reddish-orange background ("deserts"), and 3) the polar caps.

The nature of the dark maria is not very well understood because, apart from showing fairly regular seasonal changes, they also are subject to erratic variations that make the hypothesis of Earth-type vegetation somewhat doubtful. Moreover, the infrared spectrum does not have characteristic absorption bands of chlorophyll.[12] The presence of organic material in these regions, however, has been reported by Sinton,[31] who observed C-H vibration absorptions in the 3.5 μ region which were absent in the reflection spectrum of the desert areas.

Another factor requiring explanation regarding the dark areas is the higher temperature observed by Sinton and Strong[24] which indicates that on an average they are 8°K warmer than the bright areas. If the "maria" were areas of Earth-type vegetation, one would expect a higher water vapor content than in the desert regions at the same temperature. This in turn would mean that the measurements from Earth in the 8 to 12 μ window would be contaminated by water vapor emission from the atmosphere, and one would expect to see a lower temperature in the darker regions (cf., observations by Tiros in the 8 to 12 μ region of central Africa and the Sahara).[32] The observed higher temperature in the dark areas of Mars is, however, in agreement with the lower albedo of these regions if interpreted as a strictly surface phenomena.

The probability of the existence of life on Mars is extremely controversial and is beyond the scope of this review on planetary atmospheres, but it will be desirable to see Sinton's observations of C-H bands on Mars repeated and probably substantiated by infared spectroscopic measurements of the terrestrial "dark and bright" areas from an Earth satellite.

The large orange areas of Mars are responsible for the reddish color of the planet seen by the naked eye. According to Kuiper,[12] they are composed of felsite rhyolite. Dollfus,[18] from polarization studies, concludes that these areas are predominately limolite, which is pure dehydrated iron oxide. The exact nature is therefore still in debate. De Vaucouleurs[23] has treated this subject in great detail, and no later observational results have been obtained which would modify this discussion.

The polar caps are rather reliably known to be a thin layer of H_2O frost deposited on the surface. Both Kuiper[12] and Dollfus[18] agree on their nature and estimate their thickness as varying between 1 and 5 cm. These caps are observed to expand to lower latitudes during fall and winter and to recede to high latitudes during spring and summer. The evaporation of the polar caps in summer and the almost immediate transport of the water vapor towards the winter pole across the equator are difficult to understand in view of the fact that in Earth's atmosphere the hemisphere to hemisphere mixing is extremely small. De Vaucouleurs[15] has estimated the average speed of this humidity wave across the planet to be $\sim$45 km/day. This pole-to-pole circulation has no parallel on Earth and perhaps can be understood only by the greater length of the Martian year and the increased temperature difference between the cold and warm poles.[15]

Blue Haze Layer

Photographs of Mars taken through filters at wavelengths less than 4500 Å indicate no surface features, and it is therefore believed to be covered with a haze layer known as "blue haze," absorbing at these wavelengths. Sometimes during the opposition, however, it clears over certain regions, and surface features become observable in the blue. These "blue clearings" last usually a few days. Slipher[33] has reported a clearing during the opposition of 1954 lasting for two months. Conflicting arguments have been extended to explain the nature of the blue haze, but an explanation that satisfies all the observed features is yet to come.

Kuiper has suggested that the haze is made of ice crystals of size 0.3 to 0.4 μ located at a height of 6 to 10 km. But, because of the fact that the albedo of Mars decreases very sharply in the blue, the presence of material that absorbs radiation of wavelength less than 4500 Å is required. As water drops or ice crystals are both transparent in the blue, Goody[19] suggests that the particles may be composed of dust of the desired absorbing properties. However, if the particles are nonvolatile, then the explanation of the blue clearing becomes still more difficult. Sharanov,[34] on the other hand, proposes that the blue haze is an optical phenomenon readily explainable by particle and Rayleigh scattering in the lower atmosphere of Mars.

Urey and Brewer[35] have pointed out that CO_2^+, N_2^+, and CO^+ ions absorb in the blue and ultraviolet but fluoresce in the longer wavelengths. The clearings at the opposition then can be explained by the diminution of the solar particle radiation arriving at Mars due to the deviating effect of Earth's magnetic field. Sagan,[36] however, recently has given estimates of the solar particle flux at the distance of Mars and its variation due to Earth's magnetic field and concludes that such mechanism is infeasible.

As mentioned in the previous section, Ohring[6] has suggested that a frozen CO_2 cloud layer at an altitude of $\sim$30 km in the Martian atmosphere may account for the "blue haze" phenomenon. If such were the case, the haze layer would obscure the surface at *all* visual wavelengths,[37] and the absorption would not be confined only in the blue.

Also, if the haze layer were situated at a high level in the Martian atmosphere as suggested by the hypothesis of ions absorption or stratospheric "dry ice" layer, then the apparent diameter of the planet in the blue and in the red should be considerably different. The very recent measurements of Dollfus[38] indicate a difference of less than 0.1% in the two diameters, thus implying that the level of the haze layer on

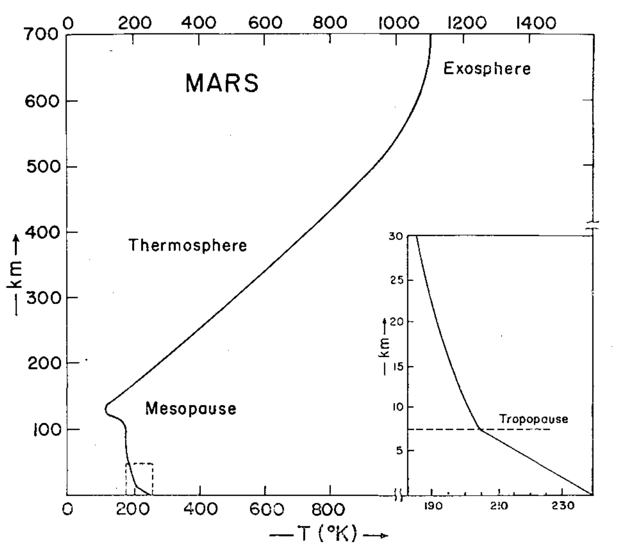

Fig. 4 Probable vertical temperature structure of the Martian atmosphere (after Arking, 1962[4] and Chamberlain, 1962[5])

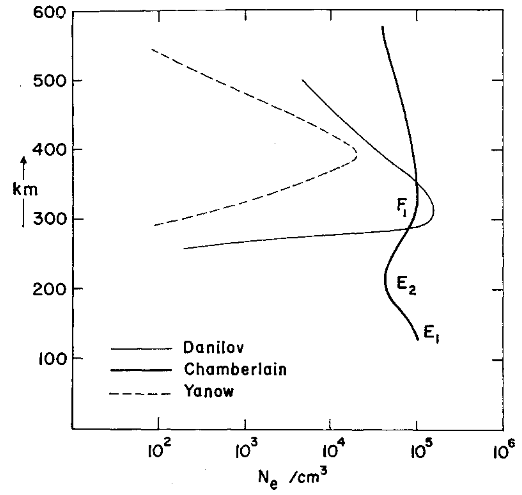

Fig. 5 Ionospheric models for Mars (electron density vs altitude)

Mars may not be at an altitude higher than 10 km. An interesting explanation of the "blue haze" recently has been proposed by Warneck and Marmo.[39] It has been shown that a trace amount of NO_2 in the atmosphere of Mars can provide the observed opacity in the blue wavelength region. A more detailed and careful examination of this possibility, however, is desirable.

Clouds

Several types of thin clouds frequently have been observed on Mars. They can be divided into three types:

1) Blue clouds are visible only in the blue. They are in patches and are seen near the poles and near the terminator. Polarization measurements indicate their particle size to be approximately 0.1 μ[18] and they occur probably at altitudes less than 100 km (see discussion of blue haze). According to Goody,[19] they may be composed of ice crystals formed on the nuclei fed by the haze layer, whereas others believe them to be of the same material as the blue haze.

2) White clouds are visible in both the blue and yellow light. The polarization studies of these clouds suggest their nature to be the same as ice crystal clouds of a size of approximately 1 μ.[18] Being composed of bigger particles, they probably lie at altitudes less than that of blue clouds.

3) Yellow clouds, visible only in yellow, are very rare and variable in size. They have been seen drifting several hundred miles across the planet at a velocity of 60 km/hr.[15] According to Goody,[19] they are composed of the same blue absorbing material as the blue haze. The reason why they are not visible in the blue is that they lie above the blue haze, and as the haze is optically thick in the blue and thin in the yellow and red, an overlying cloud of the same material will be seen only in the yellow and red. Sinton and Strong[24] have measured the temperature at the top of these yellow clouds and find a value $\sim$25°K less than the probable ground temperature. Hess,[21] assuming an adiabatic temperature gradient of -3.7°K/km, estimates the height of yellow clouds to be $\sim$6 to 7 km.

Upper Atmosphere and Ionosphere of Mars

Assuming an atmospheric composition of 98% N_2 and 2% CO_2 and a Martian stratospheric temperature of 134°K,[19] Chamberlain[5] recently has computed the mesospheric cooling and thermospheric heating for Mars. With the assumed model atmosphere and from the considerations of CO_2 dissociation into CO and O and the consequent CO cooling at the mesopause, Chamberlain deduces the height of the mesopause as $\sim$130 km at a temperature of 76°K. The considerable CO cooling at the mesopause level acts as "an effective thermostat, keeping the temperature at the exospheric or escape level (1500 km) from exceeding 1100°K."[5] The lifetime of oxygen on Mars according to Eq. (1) turns out to be 10^9 yr,[5] implying its probable retention by the planet. Urey[10] previously had argued that, because of the greater height of the escape level on Mars, thermal conduction to lower layers must be less effective, giving the exospheric temperature of the order of 2000°K. This meant that all gases with atomic or molecular weight <20 certainly would escape from the planet in time $<10^9$ yr. In this exospheric model, however, CO cooling at the mesopause was neglected.

It is difficult to say how rigidly the 1100°K figure applies to Mars in actuality. In the case of Earth's exosphere, it recently has been shown that, though the mean temperature is of the order of 1400°K, the diurnal variations have an amplitude of as much as 500°K.[40] Moreover, short- and long-period variations in the temperature value, highly correlated with the solar activity, also have been observed.[40,41] The daytime exospheric temperature, at solar cycle maximum, may attain a value as high as $\sim$2200°K.[40]

For the considerations of the gravitational escape of gases, the maximum temperature is more relevant. Hence, in the case of Mars, even if one assumes the average exospheric temperature to be 1100°K, the escape of gases will be governed by the day-side maximum temperatures attained during periods of high solar activity. Certainly more oxygen would escape if the exospheric temperature of Mars occasionally rose to 1500°K (even for a total time of $\frac{1}{10}$ of planetary history) than if it remained steady at 1100°K. The retention of oxygen by Mars is less probable than estimated by Chamberlain.

Figure 4 gives a rough picture of the vertical temperature profile of the Martian atmosphere. The temperatures up to an altitude of 100 km are based on the results obtained by Arking,[4] whereas for above this altitude the upper atmospheric model developed by Chamberlain has been used.

Danilov,[42] Yanow,[43] and Chamberlain[5] have computed the probable electron densities in the ionosphere of Mars, shown in Fig. 5. Danilov has considered an atmosphere composed of N_2^+, NO^+, O^+, and O_2^+. The curve drawn after Chamberlain in Fig. 5 is an upper limit to the possible electron densities but is extremely useful, because it shows the possibility of a secondary maximum to be as low as 130 km. As pointed out by Kellogg and Sagan,[37] Yanow has not considered the importance of electron attachment, and his curve given here shows only O^+ density as derived for 98% N_2 and 2% CO_2 atmosphere, considering a three-body recombination of O with CO or N.

The forementioned ionospheric models are based on the assumption that only the solar ultraviolet and x rays are the ionizing agents in the upper atmosphere. If Mars has a weak magnetic field, the ionization by solar proton flux could be very significant. Also, as the orbit of Mars is near the asteroid belt, a heavier stream of meteoric dust particles is expected to be entering the planet. There is some evidence that Earth's *E* region is partly maintained by the energy supplied by such meteor streams,[44] and, if the accretion rate of interplanetary dust at Mars is really higher than at Earth, a more important effect on the ionosphere of Mars can be expected. These possibilities are, however, highly hypothetical and have been mentioned only to emphasize the tentative nature of the curves given in Fig. 5.

Venus

Venus is our nearest planetary neighbor and, after the sun and moon, the brightest object in the sky. It therefore has attracted the attention of man since the beginning of civilization. Despite the great interest, very little is known about the atmosphere of this planet, especially when compared with the information available about Mars. The main reason for this deficiency is that Venus is covered with a layer of white

Table 4 Astronomical data for Venus

	Mass	Rad.	Distance, a.u.	Density	Albedo	T_e, °K	g, cm/sec^2
Earth	1	1	1	5.5	0.39	245	980
Venus	0.81	0.97	0.72	4.8	0.73[a]	235	842

[a] This new value of albedo (cf., 0.76 given by Kuiper[12] and hitherto generally accepted) takes into account the lower albedo of Venus in the infrared recently reported by Sinton.[45]

clouds, and the surface remains permanently invisible. Observations have been made, however, in the infrared and radio-frequency regions, and new information regarding the composition and temperature distribution in the Venus atmosphere has been obtained in the past few years. These new results of temperature and pressure at the various levels in the Venus atmosphere have forced a complete revision of ideas regarding the atmospheric structure of this planet. The author will try to summarize the present-day knowledge regarding the Venus atmosphere by giving the physical constants of the planet (Table 4) and reviewing the various hypotheses relative to the structure of the Venus atmosphere.

Composition

From the analysis of reflected solar spectrum, the only atmospheric constituent so far detected in Venus is CO_2.[46] Its abundance above the effective "reflecting level" of the 8000-Å photon has been estimated by several workers.[47,48] Until recently, the amount of CO_2 above the cloud level was accepted to be 1000 m-atm,[10] which made up 95% of the total atmosphere above the clouds. Recent reinterpretations of the old spectra of Venus by Spinrad,[2] however, give a CO_2/atmosphere ratio by volume of only 5%, which for a cloud top pressure of 90 *mb* (see discussion on pressure) will correspond to only 40 m-atm of CO_2 above the clouds.

Attempts also have been made to detect water vapor in the atmosphere of Venus. From high altitude balloon measurements, Strong[49] in 1960 obtained a value of 2×10^{-3} g/cm^2 for water vapor above the cloud. The interpretation of these measurements, however, must be questioned in the light of subsequent observations,[50] which indicate a comparable amount of water vapor in Earth's stratosphere. If, however, the clouds of Venus are composed of water drops or of ice crystals,[51,52] a large abundance of water vapor can be expected below the clouds. The amount of water vapor above will be limited by the saturated vapor pressure at the temperature of the cloud tops. Recently Spinrad[2] also has looked for water vapor bands but has failed to detect any.

No other constituent in the Venus atmosphere has been observed spectroscopically, except that recent high resolution measurements of Sinton[45] give a value of 4 cm-atm for CO above the relevant reflecting level of the 2.0 μ photon, but Kuiper,[53] by similar measurements, has not been able to observe any CO bands. Upper limits have been placed on the abundance of other gases that could be expected in the planet's atmosphere.[12] Interpreting the results of Dunham,[16] Urey[10] estimates the amount of oxygen above the clouds to be $\sim$80 m-atm. Again, according to Urey,[10] because of the presence of carbon dioxide, CH_4 and NH_3 cannot be expected to be important constituents of the Venus atmosphere. Also, since all the oxides of nitrogen are unstable with respect to oxygen and nitrogen, they cannot be present in large quantities in any planetary atmosphere.[10] Because it lacks absorption bands in the visible and in the near infrared, nitrogen cannot be detected by Earth-bound observations. Because of its cosmic abundance, however, the presence of N_2 in Venus is most likely, and, in order to account for the pressure, it is generally assumed that the rest of the atmosphere is composed of nitrogen.

Based on Spinrad's estimate of CO_2 mixing ratio, the following composition will be adopted for the Venus atmosphere: $N_2 \sim 95\%$, $CO_2 \sim 5\%$, $H_2O \sim$ indefinite, and $O_2 < 80$ m-atm above the clouds.

Pressure

There are two rather reliably determined pressure points in the Venus atmosphere:

1) Dollfus[54] has observed a difference in polarization in the red and green and has interpreted this observation as the result of molecular scattering in an 800-m-thick atmosphere the cloud top. For the slightly lower gravity of Venus, this corresponds to a pressure of 90 *mb* at the cloud top.

2) De Vaucouleurs and Menzel,[55] by their observations of the occultation of Regulus by Venus, estimate a pressure of 2.6×10^{-3} *mb* at an altitude of $\sim$70 km above the top of the clouds.

Temperature

The temperature measurements are more numerous, but the altitudes in the atmosphere to which they refer are very uncertain. Sinton and Strong[56] have confirmed the initial 8 to 13 μ radiometric temperature measurements[57] and found a value of 235 $\pm$ 10°K. This temperature should originate from a level where the CO_2, if it is the principal radiator in this spectral region, would become optically thick as seen from the top of the atmosphere. The 9.4 and 10.4 μ absorption bands of CO_2 are extremely weak, and, to make CO_2 radiate effectively in this region of the spectrum at relevant pressure and temperature, one must have a depth of at least 400 m-atm. If there are only 40 m-atm of CO_2 above the cloud, the origin of the temperature 235°K is then probably at the cloud top.

From an analysis of the intensity distribution within the CO_2-8000°K-vibration-rotation band levels, Chamberlain and Kuiper[58] estimate a temperature of 285 $\pm$ 9°K. The same photographic plates of the Venus spectra obtained in 1932 at Mt. Wilson were recently re-analyzed by Spinrad, and he obtained a rotational temperature varying between 250° and 440°K from plate to plate, i.e., from day to day, which also corresponded to different values of pressure—the low temperatures corresponding to lower pressures, and higher temperature presumably originating from a pressure level up to 6 atm. Spinrad has interpreted these results as probable breaks in the cloud and suggested that the higher temperatures and pressures probably refer to levels far below the visible cloud layer.

De Vaucouleurs and Menzel,[55] while observing the occultation of Regulus by Venus, have deduced fading rate due to differential refraction in the upper atmosphere of Venus and have obtained a scale height $H = kT/mg = 6.8$ km and for its first derivative $(1/H) \times (\delta H/\delta z)$, a value of 1%/km. The authors, assuming a mean molar mass of 42.5, estimated a temperature of 297°K and a temperature gradient of 3°K/km at a height of 70 km above the cloud layer. If, however, the assumed atmosphere is 95% N_2, then this assumption of $M = 42.5$ is too high. For $M = 29$ (for an N_2-CO_2 atmosphere as proposed by Spinrad), one obtains a temperature of $\sim$203°K and a positive temperature gradient of 2°K/km, indicating that the minimum temperature at the mesopause of Venus is probably $\sim$200°K. From the consideration of the change in pressure from 90 *mb* at cloud level to 2.6×10^{-3} *mb* at 70 km above the clouds, one derives a mean scale height of 6.7 km. For a mean molecular weight of 29, this will correspond to a mean temperature of 207°K. This will imply that the temperature in the stratosphere of Venus, for the most part, remains $\sim$200°K. If the cloud top pressure is $\sim$90 *mb* and if 40 m-atm of CO_2 is the only absorbing gas above it, the

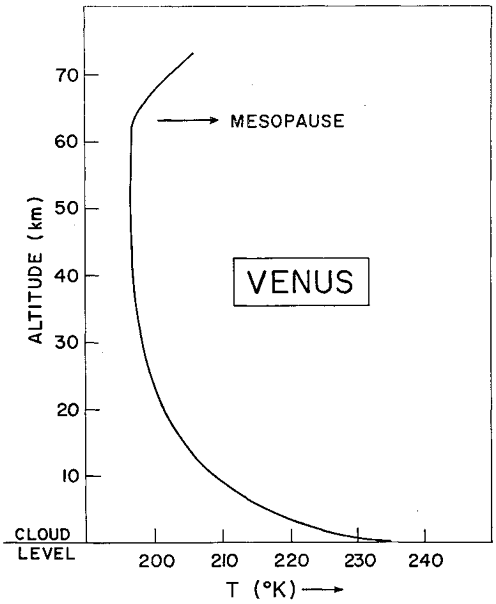

Fig. 6 Radiative temperature profile of the Venus atmosphere above the clouds

problem is very similar to that of the atmosphere of Mars, the cloud level here corresponding to the surface of Mars. The vertical temperature profile in an atmosphere in radiative equilibrium and having a given optical thickness can be estimated approximately from the following simple relation:

$$T_{(z)}^4 = T_e^4(\tfrac{1}{2} + \tfrac{3}{4}\tau_z)$$

where $T_{(z)}$ is the temperature at the level z, T_e is the effective blackbody temperature of the planet, and τ_z is the optical thickness of the atmosphere above the level z:

$$\tau_z = \int_z^\infty \kappa\rho dz$$

Knowing the temperature at the cloud level $T_{(z=0)}$, one can estimate the optical thickness of the atmosphere above the clouds. For Venus, $T_e \sim 235$°K and $T_{(z=0)}$ at cloud level also (by pure accident) ~ 235°K; $\tau_{(z=0)}$ then is ~ 0.7. This implies that the mean transmission in the infrared of the atmosphere above the clouds is $e^{-0.7} \sim 50\%$. Now τ_z, at any other altitude above, varies as

$$\tau_z = \tau_{(z=0)} \exp\left(-\int_{z=0}^{z} \frac{dz}{H}\right)$$

where H is the scale height $\sim$7 km, and so one can compute the temperature profile for the atmosphere above the cloud level. The temperature attains an asymptotic value of 197°K at $\sim$20 km above the cloud and, because of photodissociative and photoionizing reactions, probably starts increasing again at $\sim$60 km. Figure 6 shows the probable temperature profile of the atmosphere of Venus above the clouds if it is in purely radiative equilibrium.

It is interesting to compare this result with that obtained by Mintz,[59] who assumed a much greater amount of CO_2 above the clouds and thereby obtained a different temperature profile. The extent of increase in temperature in the ionosphere is not very well known and requires a careful treatment, because cooling processes due to CO, CO_2, and O_2 largely control the temperature at the mesopause.[5]

The structure of the lower atmosphere of Venus is very poorly understood. Urey[10] in 1959 had estimated a ground temperature for Venus of the order of 320°K, but recent measurements of the intensity of the radiation emitted by the planet in the centimeter wavelength show that it corresponds to thermal radiation of temperature of $\sim$600°K.[1] Since radiation in the decimeter region probably passes through the atmosphere and clouds of Venus, it generally is assumed that the measured temperature refers to the surface of Venus. Also, Sagan[52] has shown that the observed temperature spectrum (Fig. 7) cannot be interpreted as synchrotron or cyclotron radiation. The interpretation of these measurements is as yet one of the most important problems in the physics of planetary atmospheres.

Though the observational points are still very few and some of the results are fraught with uncertainties (e.g., variation of measured radio temperatures with phase angle, period of rotation of the planet, tilt of the axis, etc.), several atmospheric models already have been derived in order to explain the high radio-brightness temperatures. Kellogg and Sagan[37] recently have published an excellent review of the proposed model atmospheres for Venus, and therefore they will be described only very briefly, with special emphasis on more recent results.

The temperature measurements shown in Fig. 6 have been made mostly at inferior conjunction of Venus. The high temperature of $\sim$580°K therefore refers to the night side of the planet. As the phase angle decreases, the radio measurements become more and more difficult, i.e., the signal to noise ratio becomes smaller and smaller, making the measurements very uncertain. Mayer[1] has recently summarized the measured temperatures as a function of phase angle. There is some indication of higher than 600°K temperatures for smaller phase angles, but because of the uncertainty of $\sim\pm 100$°K in each temperature value, the existence of a phase effect cannot be established. More recently, from the 1961 inferior conjunction measurements, Mayer et al.[60] and Kuzmin and Salomonovich[61] report different magnitudes of the phase effect for different wavelengths, which has made the whole problem still more confusing. The best estimates[60] at present put the dayside temperature at $\sim$750°K, but unless more observations confirm the phase effect, this value should be taken as highly tentative.

Radar Observations

Because of the complete cloud cover, no sharp features are distinguishable to determine the rotation period of Venus by optical measurements. Radar techniques recently have been employed, but the results are very controversial. Victor and Stevens[62] report a period of rotation of 225 days, which is also the period of orbital revolution of Venus, thereby indicating a synchronous rotation of the planet. According to Kotelnikov,[63] however, Venus is rotating much faster, and the period has been estimated to be between 9 and 11 days.

There is also some indication[60] that the minimum radio brightness temperature is not observed at the maximum phase angle, and instead there is a time lag of several days; Kuzmin[61] and Opik[3] therefore argue that the planet may not have a synchronous rotation. If, however, the high radio-brightness temperatures refer to the surface of the planet and

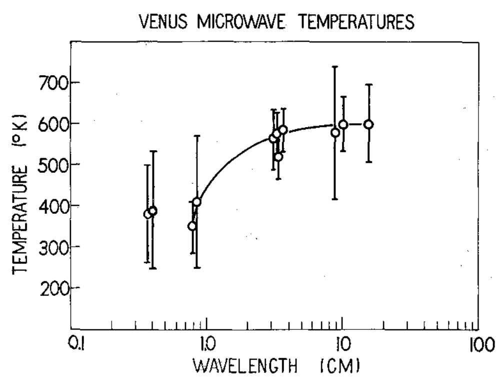

Fig. 7 Observed microwave brightness temperatures of Venus at different wavelengths (after Mayer, 1961[1])

the phase effect on the temperature is finally established, a "considerably" slower rotation of the planet is probable.

Despite these uncertainties and lack of comprehensive observational data, the following atmospheric models have been proposed.

Models for Venus Atmosphere

1 *Greenhouse model*

In this model, it is assumed that the 600°K radio-brightness temperature actually exists at the surface of the planet and is maintained by a very effective "greenhouse effect." The solar radiation (minus albedo) in the visible penetrates up to the surface; the planet, thereby being heated up to a temperature T_e, emits in the infrared, but because of the presence of triatomic molecules like CO_2 and possibly H_2O, which have strong absorption bands in the infrared region, most of the radiation remains trapped in the atmosphere and heats up the surface.

Sagan[52] has estimated the required degree of absorption as 99.5%, which, according to him, will be obtained by an atmosphere composed of 18 km-atm of CO_2 and 10 g/cm^2 of H_2O. From a more elaborate treatment of the problem of radiative transfer in a planetary atmosphere, Jastrow and Rasool[28] recently have pointed out that much higher atmospheric opacity will be required in order to obtain a Venus ground temperature of 600°K. As described previously, T_G (ground temperature) can be obtained from the relation

$$T_G^4 = T_e^4(1 + \tfrac{3}{4}\,\tau_0)$$

Inserting the value of 600°K for T_G and 235°K for T_e, one finds $\tau_0 = 42.5$, which corresponds to a transmission of only $e^{-42.5}$ or $10^{-18.5}$. An absorption of 99.5%, which corresponds to τ_0 of approximately 5, would raise the ground temperature to only 340°K. Thus, if the surface temperature has to be raised to 600°K by a greenhouse effect, the absorption by the atmosphere must be extremely high. Although the model atmosphere proposed by Sagan becomes considerably opaque at high temperatures and pressures, yet even at 600°K and 2 atm pressure the optical thickness in the infrared (τ_0) does not exceed 12.[28] If, therefore, the 600°K surface temperature is attained by a greenhouse effect, then apparently the actual atmosphere of Venus is much different from the model adopted here. Also, recent findings of Spinrad indicate that much less CO_2 is present in the Venus atmosphere than was believed so far.

Water vapor is one of the most effective absorbers of infrared radiation, and large amounts (>100 g/cm^2) at high temperatures certainly can provide infinite optical thickness. Martynov[64] also has proposed a model atmosphere containing large amounts of water vapor, but the absence of water vapor absorption bands in the spectra analyzed by Spinrad[2] (which probably refer to atmosphere below the clouds) makes this model difficult to accept.

If, however, extremely high pressures (~50 atm or more) are prevalent, the pressure broadening of individual absorption lines and induced dipole absorption probably will produce the required opacity even for relatively small amounts of CO_2 and H_2O.[28]

2 *Aeolosphere model*

Opik[65] contends that the radiative greenhouse effect cannot account for the surface temperature of 600°K and suggests that the blanketing must be due to dust, wind friction at the surface being the main source of energy for the high temperature. The dust probably is made of calcium and magnesium carbonates, and the atmosphere is composed mainly of CO_2 and N_2. No water vapor is necessarily present in the atmosphere. Because of the blanket of dust, there is no sunlight penetrating to the surface. According to this model, therefore, the radio temperatures, if they refer to the surface, should not show any dependence on the phase angle. The reality of higher temperatures at low phase angles as described previously is still controversial, and more measurements near the superior conjunction of Venus are needed to clarify the situation.

3 *Ionosphere model*

Lastly, there is a possibility that the high apparent temperature measured in the centimeter wavelengths region may refer to the ionosphere.[66] In this case, the surface would be at a temperature of ~300°K, whereas the ionosphere would contain a large concentration of free electrons up to a considerable depth, which would lead to a free-free transition and will account for this high temperature. The electron density for the ionospheric thickness of about 300 km (comparable to Earth) for such a model is about 10^9/cm^3. This is about 1000 times greater than the maximum density in Earth's ionosphere. It is very difficult to envisage such high electron concentration unless an ionization mechanism is found which is many orders of magnitude more effective than the solar ultraviolet and x rays. The solar proton flux has been suggested[66] as the other source, but, from its value as known at the distance of Earth and assuming that Venus has a very weak magnitude field, this would provide an ionization mechanism at the maximum of only ~30 times more effective than the ultraviolet radiation.[37]

At the moment, no known mechanism can provide such high electron densities as 10^9/cm^3 in the Venus ionosphere, and, moreover, there is another fact observed which argues against the ionospheric model. If the ionosphere is optically thick at $\lambda = 3$ cm, it certainly will be opaque at 12.5 cm, which is the wavelength used for the radar measurements. Now, as the ionosphere cannot be opaque and reflect at the same time, the electron density should be still higher (~10^{12}/cm^3) in order for the ionosphere to reflect at 12.5-cm wavelength. It also has been suggested that there is an ionospheric hole at the midnight point (or antisolar point) of Venus,[37] and, as the radar measurements have been made only at the inferior conjunction, this may account for the radar reflections. These suggestions are, however, highly speculative, and an ionosphere of such high electron density in the first place and its failure to account for the radar reflectivities make the ionospheric model for the explanation of 600°K temperature very improbable.

Apart from the high temperature observed at wavelengths >3 cm, there is another aspect of the measured temperature values (Fig. 7) which requires explanation: the apparent decrease in the brightness temperatures at wavelengths < 1 cm.

The forementioned atmospheric models give a tentative explanation of the low temperature obtained at 8 mm. In the case of the greenhouse and aeolospheric models, the radiation corresponding to 350°K as observed at 8 mm is emitted from the middle of the troposphere, whereas in the ionospheric model, in which the ionosphere is emitting at 600°K, the lower temperature refers to the surface of the planet.

Barrett[67] has shown that self-absorption of an atmosphere composed of CO_2 and H_2O at a total pressure of ~20 atm and surface temperature of 600°K will account for the observed temperature spectrum. Rasool,[22] on the other hand, attributes this sudden decrease of temperature at 8 mm to 50% attenuation of microwave radiation at this wavelength by a 2-km-thick cloud layer, assuming that the clouds of Venus were of the terrestrial type made of water with a drop size of less than 50 μ. These models were conceived before the 4-mm temperature measurements had been made, and, except for the ionosphere model, these explanations required a still lower temperature at 4-mm wavelength.

According to Fig. 7, however, the temperatures measured at 4 mm are of the same order of magnitude as at 8 mm. If the level of origin of the radiation at the two wavelengths is the same, then the forementioned interpretations become questionable. The accuracy of temperature measurements at 4 and 8 mm is, however, so low that, before drawing any definite conclusions, more observational points have to be obtained which would give a definite shape to the temperature spectrum. More measurements in the wavelength interval 0.1 to 3.0 cm will be extremely rewarding.

The confused state of knowledge regarding the atmosphere of Venus as outlined previously probably will not last very long. The data relayed back by Mariner II[68] may contain useful clues regarding the physical conditions prevalent at the surface and in the atmosphere of Earth's "sister" planet, which appears to be so different.

Jupiter

Because of their considerable distances from Earth and the probable presence of a thick atmosphere, relatively little is known about the structure of the atmospheres of the major planets. Jupiter is the biggest of all the planets, with a mass ~300 times greater than Earth. As its volume exceeds Earth's by a factor of 1000, the mean density is comparatively small (1.33 g/cm^3, Earth = 5.5 g/cm^3). Being 5 times farther removed from the sun than Earth but having a comparable albedo of 0.47, purely physical considerations indicate that the temperature of the planet should be very low. With a large surface gravity (2.6 times that of Earth) and a lower temperature, one would expect the chemical composition of the atmosphere to be still primitive, containing large quantities of hydrogen and helium. An estimate on the basis of Eq. (1) gives the time of escape of hydrogen from Jupiter of the order of 10^{116} yr. This excludes the possibility of the gravitational escape of any atmospheric gas from the planet and indicates a predominantly hydrogen atmosphere. Recent theoretical and experimental evidence is, however, against this composition of the Jovian atmosphere, and the author will therefore consider the most recent observational results in more detail and attempt to understand the structure of the atmosphere of Jupiter.

Composition

The only gases spectroscopically detected are CH_4 and NH_3. Their abundances, according to Kuiper,[12] are CH_4 = 150 m-atm and NH_3 = 7 m-atm above the cloud surface. As both hydrogen and helium are undetectable by spectroscopic measurements in the visible, no direct evidence of their presence in Jupiter's atmosphere was available until very recently, when Kiess, Corliss, and Kiess[69] detected the quadrupole rotation-vibration lines of molecular hydrogen in the Jovian spectra. Zabriskie[70] analyzed these spectra to obtain a total hydrogen amount of 5.5 km-atm of molecular hydrogen above the cloud level. This, however, is in complete disagreement with the hitherto accepted atmospheric composition of Jupiter which, according to Urey,[10] is as follows:

Gas	km-atm
H_2	270
He	56
CH_4	0.15
NH_3	0.007

The hydrogen and helium abundances had been derived on the assumption that they are present in solar proportions. In that case, neon and nitrogen also will be present in small quantities. The mean molecular weight of this atmosphere would be 3.25 and the pressure at the cloud layer ~8 bars.

Very recently Opik[3] has questioned this composition because of the following reasons: Baum and Code[71] observed the occultation of σ Arietis by Jupiter. From the rate of fading of the occulted star, they calculated the density scale height $H = kt/mg \sim 8.3$ km for the atmosphere of the planet, and, from this observed value of scale height and the polychromatic radiative equilibrium temperature value of 112°K, Opik estimates the mean molecular weight of the atmosphere of Jupiter above the clouds to be 4.3 ± 0.5, which is considerably different from the value obtained for the previously quoted abundances. Opik therefore accepts the measured hydrogen abundance of 5.5 km-atm[70] and proposes the following composition for Jupiter:

Gas	Percentage
He	97.2000
H_2	2.3000
Ne	0.3900
CH_4	0.0630
Ar	0.0642
NH_3	0.0029

which will give a mean molecular weight of 4.02, in good agreement with the occultation value. Opik also argues against the presence of free nitrogen in the atmosphere which will combine immediately with hydrogen to form NH_3. CO_2 also will be absent, having been reduced to CH_4 and H_2O. Water would remain beneath the clouds and therefore would be undetectable from Earth. If this composition is accepted, the fractionation of hydrogen in the earlier history of the planet has yet to be explained.

Temperature

The effective blackbody temperature of Jupiter for a visual albedo of 0.47 is 105°K. Measurements have been made in the 8 to 12 μ window by Menzel et al.[72] which indicate a value of ~130°K. This measurement may not refer to the cloud top level of Jupiter, as both NH_3 and CH_4 emit in the far infrared, and the measured radiation probably may be weighted heavily with atmospheric emission.[73] The measurements at radio frequencies, however, give a very different result. Table 5 summarizes the up-to-date observational results at different frequencies.[1]

It is believed that the radiation observed at ~3 cm wavelength is thermal, but the level to which it refers is not known because it is probably heavily weighted by the pressure-broadened 1.28-cm line of NH_3.[73] The highly intense radiation observed at wavelength >10 cm is probably nonthermal, originating from the Van Allen-type belts around Jupiter.[74] A brief discussion of these radiation belts will be given in a later section.

Clouds

Telescopic view of Jupiter shows banded cloud structure. The period of rotation being very small ($9^h\ 55^m\ 28^s$), the banded cloud layers can be explained easily. The nature of these clouds and the color effects observed in these cloud belts are, however, still controversial and recently have been discussed in detail by Newburn.[73] The clouds probably are made of NH_3, which is quite comprehensible in view of the prevalent low temperatures and high pressure. The observed

Table 5 Observed radio brightness temperatures of Jupiter

Wavelength, cm	Temperature, °K
3.03	171
3.17	173
3.36	189
3.75	210
10.20	640
21.00	2,500
22.00	3,000
31.00	5,500
68.00	70,000

motion of the clouds, however, poses an interesting problem of meteorology. The equatorial clouds have a period of rotation ~5 min less than those at higher latitudes. Kuiper[12] has given a tentative explanation of this difference in rotation period by assuming a gaseous ring around the planet which would accelerate the equatorial clouds. The problem is far from being resolved and probably is related to the possible differences in the tropopause heights at the equator and at the poles.[75]

Red Spot

Another problem of interest in the atmosphere of Jupiter is the great red spot, which is ~40,000 km in length, ~13,000 km in width, and was first observed in 1831. This spot reached its highest intensity in 1880, when its color became pink. Since then, both visibility and color have waxed and waned. Moreover, it is not attached rigidly to the surface, as it has been observed to oscillate at random. The most common explanation is that of a large meteorite floating on liquid or in a heavy atmosphere, but the changes in color and intensity cannot be accounted for adequately by this hypothesis. Recently, Hide[76] has advanced an explanation in which the surface of the planet in that region would be a plateau only a few kilometers high. The hydrodynamic theory of circulation suggests that, because Jupiter rotates so rapidly, the effect of a shallow topographical feature on the general circulation of the atmosphere will be attenuated very slowly with height. Thus, the feature will make its presence manifest at the level of the cloud. This explanation seems quite plausible, except that it is difficult to envisage the presence of only one topographical feature of this size and the absence of other shallow topographical features creating other spots that have not been observed. Moreover why should it have appeared suddenly in 1831?

Radiation Belts

Considerable interest has developed in the origin of centimetric radiation from Jupiter (Table 5) which corresponds to extremely high temperatures. Earlier investigations by Field[74] led to the possibility that the observed decimetric radiation was cyclotron and was originating from electrons trapped in a 1200-gauss polar magnetic field. Recently Roberts and Huguenin[77] have observed variation of percentage polarization with solar activity, thus eliminating cyclotron radiation as the mechanism responsible for the Jovian decimetric emissions. These measurements are more consistent with synchrotron radiation originating from highly relativistic electrons trapped in the Van Allen-type belts of Jupiter. These radiation belts presumably are highly populated with energetic particles to a distance of 3 Jupiter radii,[77] and the surface magnetic field is of the order of 50 gauss.

Another source of radiation has been observed in the decameter wavelength region and has been found to be very erratic. The spectrum of this radiation extends from 10 to 21 m, with a maximum around 15.8 m.[78] It has been suggested that this radiation may be coming from the atmosphere or from the surface but is modified by the ionosphere in the presence of magnetic field of about 7 gauss.[79] Warwick,[80] however, explains this radiation in terms of precipitation of fast electrons out of Jupiter's radiation belts and down to the surface of the planet along dipole lines of force. This explanation also will be consistent with the synchrotron model of the decimeter radiation.

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April 23–25	AIAA-ASME Hypersonic Ramjets Conference[4]	White Oak, Md.	Past
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