1962ApJ...136..582C

UPPER ATMOSPHERES OF THE PLANETS*

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Received February 10, 1962

ABSTRACT

Most of our knowledge of the earth's upper atmosphere is obtained from experiments *in situ* or remote observations that offer fairly direct interpretations. For the other planets, neither approach is as yet possible; so indirect, theoretical procedures are required to construct models of their atmospheres. Herein the wide variety of basic physical processes governing the structure of a planetary high atmosphere are set forth, with a view toward obtaining "deductive models"—i e , models derived theoretically when only the chemical composition and temperature of the lower atmosphere and the incident solar flux are specified (in addition to various physical and astronomical constants).

The procedures are applied to Mars, whose lower and middle atmosphere is already partially understood from earlier work, notably that of R. Goody. The uncertainties involved at various stages in the construction of such a model are emphasized, as are the major differences in physical processes and atmospheric characteristics between Mars and the earth.

The most significant result is that the CO that must be in the upper atmosphere of Mars should serve as an effective thermostat, keeping the temperature at the escape level (1500 km) from exceeding about 1100° K. This is cool enough for Mars to retain atomic oxygen. Were it not for CO cooling, the upper atmosphere of Mars would be so extensive and form such an effective thermal insulation between the upper ionosphere and the heat sink at the mesopause that the temperature would exceed 2000° K. The lifetime for the escape of oxygen is estimated as 10^9 years.

The mesopause is determined by CO_2 radiation. Near the mesopause CO_2 should become dissociated and the free O atoms form a thin layer of O_2; this effect has no analogy on earth. The model ionosphere has considerably smaller densities of ionization than comparable regions in the earth's atmosphere The E region is split into two distinct portions, with X-rays forming the higher one (E_2) and ultraviolet light ionizing O_2 near the mesopause (E_1) The Martian analogue to the terrestrial F_2 region may not develop a very high electron density, and the entire ionosphere should be depleted at night.

I. INTRODUCTION

Studies of an astrophysical nature have made unique contributions to our understanding of the earth's upper atmosphere, a topic generally regarded as more a part of geophysics than of astronomy. In much the same fashion, astrophysical investigation should be able to clarify many problems pertaining to the high atmospheres of other planets. The astrophysical approach is distinct from the modern-day geophysical approach, in that the fundamental data employed are only those that can be measured from well outside the atmosphere or derived from other such data. In developing this deductive technique, one is guided by current understanding of the earth's atmosphere. The numerical results are bound to be rather provisional, but they should nevertheless be sufficiently accurate to serve as a guide in planning observations and experiments from orbiting astronomical observatories and deep-space probes. The procedure, simply stated, will be to review the principal effects thought to influence the temperature, density, and ionization of the earth's upper atmosphere and then to apply each of these considerations in turn to the atmosphere of Mars.

Our objective is an atmospheric model—values of temperature, density, chemical composition, and density of free electrons against height. These parameters are interdependent, the main difficulty being the temperature dependence on density. Figure 1 compares the temperature profile of the earth's atmosphere with that of the model for

* Helen B Warner Lecture presented to the American Astronomical Society, meeting jointly with Section D of the American Association for the Advancement of Science in Denver, December 27, 1961.

† Alfred P. Sloan Research Fellow, 1961–1963.

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Mars. Figure 2 gives a similar comparison of densities. Other major conclusions regarding Mars are included in Table 1.

For discussion it is convenient to divide the atmosphere into different regions that emphasize the different processes important locally in the heat budget. This division and its terminology are illustrated in Figure 1.

In Section II we review in a general way the properties of an atmosphere up to the region near its *mesopause*—or high-altitude temperature minimum. The complications that go into determining this height and the temperature there are considered in Section III. Then in Section IV we treat the *thermosphere*—or region of increasing temperature with height.

II. THE LOWER AND MIDDLE ATMOSPHERE

a) Troposphere

On earth the temperature has a height profile, as shown in Figure 1. In the atmosphere near the surface—the so-called *troposphere*—the temperature would decrease upward with very nearly the adiabatic lapse rate for a diatomic gas, were it not for the presence

TABLE 1

PRINCIPAL HEIGHTS OF INTEREST IN UPPER ATMOSPHERE OF MARS

Height z (km)	Temp T (° K)	Total Density N (particles/cm^3)	Remarks
114	135	1×10^{15}	Level of vibrational relaxation
124	98	3×10^{14}	CO_2 dissociation (photochemical equilibrium)
130	76	1×10^{14}	Mesopause; E_1 region, $N_e<10^5$ (noon)
200	375	5×10^{11}	E_2 region, $N_e<5\times10^4$ (noon)
320	800	2×10^{10}	F_1 region, $N_e\approx10^5$ (noon)
1500	1100	1×10^7	Critical level

of water vapor, which releases latent heat as it condenses at high altitude. For Mars the troposphere is also very likely governed by hydrostatic equilibrium and convective transport of heat. The adiabatic lapse rate is linear with height and is

$$\frac{dT}{dz} = -\frac{\gamma-1}{\gamma}\frac{gM}{k}. \tag{1}$$

Here γ is the ratio of specific heats, g the acceleration of gravity, M the mean mass per particle, and k Boltzmann's constant. If the atmosphere of Mars is predominantly dry N_2, this adiabatic lapse rate is about $-3.7°$ K/km.

An albedo of 15 per cent gives a mean planetary temperature, as computed for simple radiative equilibrium, of 217° K. While solar radiation is absorbed mostly at the surface, the planetary emission in the far infrared arises from an effective layer somewhere in the troposphere or even the lower mesosphere. Goody (1957; also see Hess 1948) suggests as a model Martian troposphere a mean temperature near the ground of 270° K, a ground pressure of 85 mb (as indicated by optical scattering), and a lapse rate of $-3.7°$ K/km. The *scale height* (or distance over which the pressure in an isothermal region decreases by the exponential factor e) at the surface of Mars is then $H \equiv kT/Mg = 21$ km, or about 2.5 times the terrestrial value. The density above 30 km is higher on Mars than on earth. The upper atmosphere of Mars is thus far more extensive than

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the earth's, and various effects to be discussed presently, such as the formation of ionospheric layers, will occur at greater altitudes than do the corresponding effects on earth.

b) Mesosphere

There is considerable doubt that Mars has a "hot" *mesosphere*—i.e., a middle atmosphere wherein the temperature increases to a peak, owing to absorption of solar radiation, and decreases again, nearly adiabatically, above the peak. On earth the mesospheric heating is due mostly to the absorption of near-ultraviolet sunlight by the Hartley and Huggins bands of O_3. In the middle atmosphere this O_3 is formed in turn by association of $O + O_2$, the O being produced by photodissociation of O_2 in the weakly

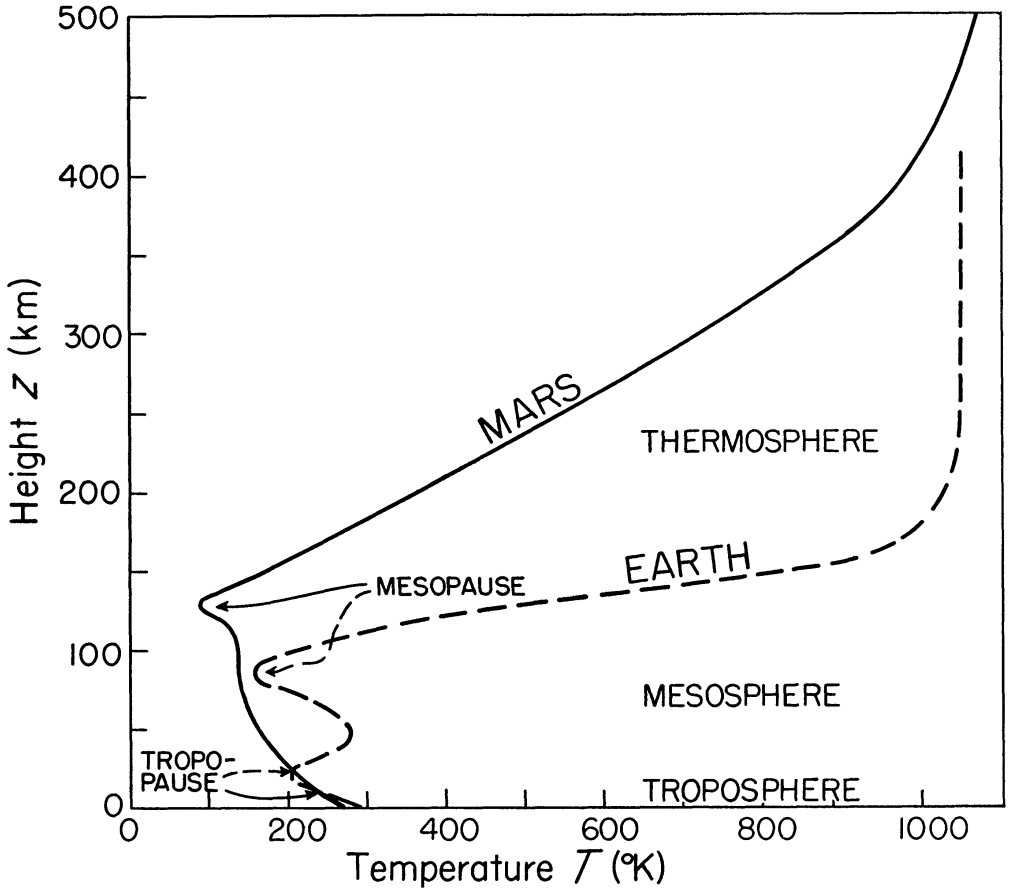


FIG. 1.—Temperature profiles of earth and Mars. The earth's thermosphere is adopted from Bates (1959), and the mesopause and lower regions are based on models given by Nicolet (1960). The Martian atmosphere up to 100 km is from Goody (1957). *Solid line*, Mars; *dashed line*, earth.

absorbing Herzberg continuum. On Mars there is not more than a trace of O_2 (less than 0.15 per cent of that on earth, according to Dunham 1949), and O_3, accordingly, is also extremely rare. The ultraviolet O_3 bands are so strongly absorbing, however, that Goody (1957) suggests there could still be sufficient O_3 on Mars to be important to the atmospheric heat balance. If present, it would at least modify the *tropopause* temperature and alter its height.

It appears likely, however, that the Martian troposphere merges into what might be called a very extended *stratosphere*. For earth this term is sometimes used to refer to the thin region immediately above the convective troposphere, where the lapse rate is more

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gradual than the adiabatic one but where mesospheric heating is not noticeable.[1] In this region the temperature is controlled by the transfer of infrared radiation, which becomes the dominant mechanism at the height where the radiative-temperature gradient becomes stable—i.e., less steep than the adiabatic gradient. Imposing the conditions that at the tropopause the gradient must thus be convectively stable and that there must be no discontinuity in the temperature between troposphere and mesosphere, Goody (1957) has computed the run of temperatures in the Martian mesosphere on two different models. In one case he considered the infrared absorption and emission to be due entirely to CO_2; in the other, the infrared-active constituent was 10^{-2} cm of precipitable H_2O.

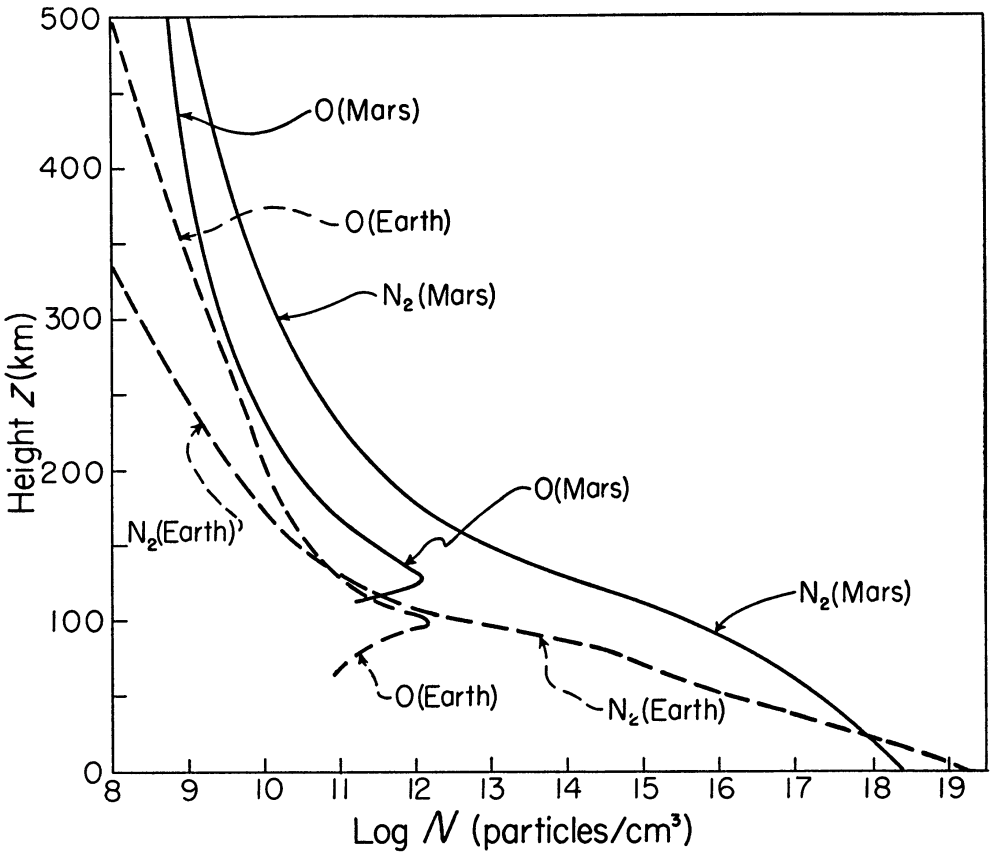


FIG. 2.—Densities in the atmospheres of earth and Mars. In computing the Martian densities, the temperature variation between the tropopause and 90 km was approximated by a linear decrease. Above the Martian mesopause (128 km) the CO density is 2×10^{-2} times the N_2 density; in the lower atmosphere CO_2 is 2×10^{-2} times the N_2 density. Diffusive separation for O begins at 120 km in the terrestrial model, at 130 km for the Martian one. *Solid lines*, Mars; *dashed lines*, earth.

The "CO_2 model" has a tropopause at 8.5 km, and the temperature decreases in the mesosphere to a constant value of 134° K above 90 km. The "H_2O model" is nearly isothermal at 153° K throughout the mesosphere, which starts at 25 km. In the following we shall adopt the "CO_2 model" to fix the boundary conditions on the mesopause and thermosphere of Mars. The temperature and density in this model are shown in Figures 1 and 2.

[1] The term *stratosphere* actually has three distinct definitions in common usage. So as not to contaminate the Martian atmosphere with ambiguities that have hindered progress on earth, I propose to delete the word entirely as being unnecessary anyway. *Mesosphere* adequately specifies the region from the tropopause to the temperature inversion at the base of the thermosphere, regardless of the extent of direct solar heating in that region.

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III. THE MESOPAUSE REGION

a) Transfer of Heat: Conduction, Radiation, Convection

Energy sources.—Above the mesosphere the temperature will again increase because of the absorption of ionizing and dissociating solar radiation and possibly other effects. The principal source of thermospheric heat is probably far-ultraviolet radiation from the sun, as shown by several studies (Bates 1956, 1959; Johnson 1958; Hunt and Van Zandt 1961). Other possibilities for augmenting this supply have been proposed from time to time, but their importance is still not demonstrated.

Nevertheless, the problem of predicting the highest thermospheric temperatures on earth from purely theoretical considerations (or, if you prefer, the problem of explaining the observed temperatures) has not yet been completely and satisfactorily solved. Bates (1956, 1959) has noted that if the terrestrial thermosphere does indeed reach temperatures approaching 1500° K, as some indirect data suggest, the solar flux *must* be supplemented at *F*-region heights.

The more popular possibilities for additional energy are conduction from the solar corona, as proposed by Chapman (1957), and particle bombardment. The more recent "hydrodynamic models" of the corona, however, predict much lower temperatures near the earth than does Chapman's static model; the heat influx to the earth would in that case be trivial. The role of particle bombardment, while quite possibly of some importance in the *F* region, cannot yet be evaluated with any confidence. Thus whether an additional source does exist and, if so, whether it would be important on Mars are matters of conjecture. Our discussion of Mars could possibly omit some important effect not yet appreciated.

Heat transport above the mesopause.—Spitzer (1949) first emphasized that, with the steep temperature gradient in the earth's thermosphere, the downward conductive flow of heat would be exceedingly important in draining energy from that region. Bates (1951) investigated the importance to the thermospheric heat budget of radiative loss by a forbidden transition of [O I] at 62 μ and by molecular vibrational transitions. A detailed discussion of the thermosphere is deferred to Section IV. For our purposes here it is necessary to state merely that the thermosphere supplies a conductive heat flux, q, flowing downward toward the mesopause.

There is a complication involved in fixing the numerical value of q at the mesopause, which depends on the total energy absorbed less the total re-radiated above that level; these processes depend on the temperature and density throughout the thermosphere. The properties of the thermosphere, in turn, are based on the properties of the mesopause. Hence the numerical solution must be obtained by an iterative process of computing the mesopause, then the entire thermosphere, and then the mesopause again. In practice, the mesopause sets in sharply and is not critically dependent on the precise value of q. Thus the second approximation is quite adequate.

Above the earth's mesosphere, solar energy is absorbed through two main processes: photoionization, occurring mainly above 120 km, and photodissociation of O_2 in the Schumann-Runge continuum shortward of 1750 A. Although the latter process absorbs sunlight mostly above 100 km, the energy dissipation into heat occurs mainly through reassociation of O atoms and will be deposited rather lower. Most of this heat is released above 85 km or so and is conducted downward toward the mesopause.

The amount of solar flux available for photodissociation is approximately (Nicolet and Mange 1954) 6×10^{11} photon/cm^2 sec. If the efficiency of conversion to heat is about 4 ev for each molecule dissociated, thermal energy is generated at an average rate over the earth of about 1 erg/cm^2 sec. Photoionization throughout the ionosphere yields a similar total heat flux (Sec. IV). The net conductive flux appearing at the mesopause will be diminished somewhat, owing to radiative losses; we shall return to this matter after considering the radiation problem at the mesopause and below.

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The temperature minimum is expected at the height where infrared radiation has dissipated all the available conductive flux from above. Below that altitude heat must be transported upward from the mesosphere by either radiation or convection.

Mesospheric radiation.—In the earth's mesosphere, radiation transport is quite important, and, in particular, the 15 μ band of CO_2 dominates the spectrum. So long as molecular collisions are frequent enough to maintain the vibrational levels of CO_2 in a Boltzmann distribution, the energy transport by this radiation may be treated by the usual techniques of radiative-transfer theory, with every absorbed quantum considered to be lost to the radiation field and with Kirchhoff's law adopted for the *source function*—i.e., the radiant energy generated in each volume element. When the pressure is so low that collisions are rare during the radiative lifetime, the gas molecules are not so efficient in taking energy out of the radiation field and transforming it into heat motions; if the kinetic temperature is then slightly perturbed by another means of heat transport, Kirchhoff's law breaks down. This problem has been treated in detail by Curtis and Goody (1956), who conclude that the role of radiation transport in governing the temperature of the mesosphere becomes greatly reduced around 75 km and above. This level is called the height of *vibrational relaxation.*

When absorption of thermal radiation is unimportant to the energy balance, the local rate of radiative energy loss may be estimated without recourse to the radiative-transfer problem. The main process to consider for the production of radiation is collisional excitation from the ground level of CO_2. If deactivating collisions are rare compared with radiative cascades, the radiation rate per unit volume is then the rate of excitation (Bates 1951),

$$R(z) = N^2(z)\, f(\mathrm{CO_2})\, \eta\, h\nu\, e^{-h\nu/kT}\,, \tag{2}$$

where N is the number density, f is the fraction by volume of CO_2 ($= 3 \times 10^{-4}$ for earth), and η is the rate coefficient for deactivation ($\approx 2.5 \times 10^{-15}$ cm^3/sec).

Location of the mesopause.—Integrating equation (2), from the mesopause at z_0 to infinity, with the (isothermal) barometric equation, $N(z) = N(z_0) \exp[-(z - z_0)/\mathrm{H}]$, for $H = kT/Mg$, gives an approximate expression for the density at which the conductive flux, q, is completely radiated away:

$$q = \tfrac{1}{2} H N^2(z_0)\, f(\mathrm{CO_2})\, \eta\, h\nu\, e^{-h\nu/kT}\,. \tag{3}$$

Even though this expression neglects any absorbed radiation or convective transport as a means of disposing of conducted heat, such terms are not likely to be more important than the radiative loss, so that equation (3) may be crudely appropriate.

The precise value of q—in practice, the heat flux a scale height or so above the mesopause—is not readily obtained because of radiative loss in the 100-km region by the [O I] emission at 62 μ. This transition is highly forbidden, and hence the populations of the atomic levels involved are in a Boltzmann distribution up to very great altitudes (Bates 1951). Near 100 km the atmosphere is optically thick; hence the heat loss is not easily evaluated, although a formal study has been published by Bauer and Wu (1954). It appears, however, that this radiation exerts only a small influence on the temperature distribution; perhaps only half the heat generated above the mesopause is lost in this fashion. With $T = 150°$ K and $q \approx 1$ erg/cm^2 sec, we obtain for the earth $N(z_0) \approx 2 \times 10^{14}$/cm^3, corresponding to about 80 km. The method used is very approximate, although it does give about the same loss rate near 80 km as do the more elaborate calculations of Curtis and Goody (1956).

Mesospheric convection.—It is not, in general, essential that the downward flow by conduction be halted at nearly the same altitude where the strong coupling between the kinetic temperature and the radiation field ceases. On the earth this coupling extends almost to the region of strong solar heating; hence there is an abrupt transition between radiative equilibrium and conductive transport. But CO_2 can be an effective cooling

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agent even when it is not in equilibrium with the kinetic temperature. If the level of vibrational relaxation were much lower, as with a greater CO_2 transition probability, we would expect a convective region above the level of vibrational relaxation and below the mesopause. This region would neither receive energy from conduction nor absorb appreciable radiation. The region just above the level of vibrational relaxation would then be sensitive to any turbulent mixing, which would impose a nearly adiabatic gradient.

b) Mars

Vibrational relaxation.—The middle atmosphere of Mars has already been treated by Goody (1957) up to about the level of vibrational relaxation. Adopting his "CO_2 model" described at the end of Section II and computing densities up to 90 km with the formula for a linear variation of scale height (e.g., see Nicolet 1960) and with the isothermal formula beyond, we find vibrational relaxation of CO_2 above 114 km. Between this height and the mesopause, we shall take the adiabatic temperature gradient, following the reasoning in the previous paragraph.

Photochemical equilibrium.—For Mars this problem is not quite as straightforward as it is for O_2 dissociation on the earth. The lower atmosphere contains about 2 per cent CO_2 by volume (Grandjean and Goody 1955), and most of the remainder is presumably N_2 (e.g., de Vaucouleurs 1954; Urey 1959). Solar radiation shortward of 1690 A dissociates CO_2 (Mahan 1960):

$$CO_2 + h\nu \rightarrow CO + O\,. \tag{4}$$

Association occurs by the three-body process,

$$CO + O + M \rightarrow CO_2 + M\,, \tag{5}$$

where M is any third body.

Since there is not more than a trace of O_2 in the lower atmosphere (Dunham 1949), it is usually assumed that O_2 dissociation and formation are negligible in the main photochemical problem. However, even with O_2 completely absent in the troposphere, the O atoms formed by process (4) may react with one another through

$$O + O + M \rightarrow O_2 + M\,. \tag{6}$$

If this is much more efficient than reaction (5), as it appears to be, the formation of O_2 would have two important (but opposite) effects. It would deplete the available O atoms, hindering CO from becoming oxidized, which would tend to lower the height of the dissociation level. Also, the O_2 thus formed would screen lower layers from the dissociating sunlight, as it does on the earth, tending to raise the height where CO_2 first breaks up. Oxygen dissociation,

$$O_2 + h\nu \rightarrow O + O\,, \tag{7}$$

is produced by wavelengths less than 1750 A, and its absorption cross-section is an order of magnitude greater than for process (4).

The photochemistry of CO_2 in the earth's atmosphere has been examined by Bates and Witherspoon (1952), who note that the rate coefficient for reaction (5) is probably 10^4 times smaller than the one for reaction (6). For Mars a solution of the four equations describing these four processes places the level of CO_2 dissociation at a total density of $N = 3 \times 10^{14}/\text{cm}^3$ and a height of 124 km. Immediately above this height, O_2 is also dissociated, and O_2 does in fact effectively screen the CO_2 below it. In the dissociation region O_2 has an abundance comparable to that of CO or CO_2; its relative abundance is higher here than anywhere else in the atmosphere. Well below the dissociation region, O_2 will vanish, unless, of course, there is more oxygen available than is necessary to oxidize CO. The total CO above the mesopause (see below) is 0.1 atm-cm.

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Diffusion and mixing.—There are probably important departures in the actual distributions from those given by photochemical equilibrium (Bates and Witherspoon 1952). The lifetime of a CO_2 molecule against photodissociation (4) in the upper Martian atmosphere is 2×10^7 sec—about 8 months. Hence CO may exist in sizable quantities well below 124 km, and CO_2 should extend well above. On the other hand, diffusive flow is also quite sluggish compared with its rate in the O_2 dissociation region on the earth, since the total density at the dissociation level is higher on Mars. The diffusion time scale, τ_{dif}, is of the order of H^2/D, where the diffusion coefficient, D, varies inversely with the density. A more sophisticated approach to the diffusion time is possible (Mange 1957) but scarcely seems justified now. At the Martian CO_2 dissociation level, τ_{dif} is of the order of 10^9 sec or 30 years.

Mixing should be very effective, however, in altering the relative abundances. The region just below the mesopause may be constantly stirred through convective overturning. Just above the mesopause, where the temperature gradient becomes convectively stable, it is not clear whether mixing is still effective. Throughout the thermospheric model in Section IV we assume that CO_2 is dissociated and that O is distributed in diffusion equilibrium; that is, we ignore mixing. But planetary circulation and possibly turbulence in the lower thermosphere could alter these abundance distributions.

Location of the mesopause.—The mesopause on Mars could be determined in the same manner as described earlier for the earth, if it is cooled mainly by radiative loss. The available heat flux at the base of the thermosphere is less than for earth, not only because of the greater distance from the sun, but also because of radiative losses at higher altitudes. We shall show in Section IV that about 65 per cent of the heat input through ionization is radiated away by CO in the thermosphere. Most of the molecular dissociation occurs in a region where CO_2 is so abundant that the radiative loss is greater than the solar energy absorbed. However, the Martian mesopause occurs in a region where vibrational relaxation exists and which may therefore be governed by convection.

An appropriate value for q at the Martian mesopause is probably 0.07 erg/cm^2 sec. Rather arbitrarily we shall place the mesopause at a height z of about 130 km, where $T = 76°$ K and $N = 1.2 \times 10^{14}$ cm^{-3}.

This temperature is about 15° below the frost point for CO_2 at the mesopause pressure. The actual mesopause temperature is uncertain, since in this region the heat transport, being affected by radiation, conduction, and convection, is most complex. Nevertheless, it seems likely that, with appropriate nuclei for sublimation, a low-temperature mesopause may give rise to a Martian analogue of the terrestrial noctilucent clouds. Further, such high-altitude ice particles, which would vaporize at times of excess heating near the mesopause, might be responsible for the elusive "blue haze."

IV. THE THERMOSPHERE

a) Heat Budget

The heat-budget equation in the thermosphere, with only photoionization and dissociation as the energy sources, may be expressed in a rather simplified fashion as

$$\frac{\partial}{\partial t}\left(\frac{n}{2} NkT\right) = \frac{\partial}{\partial z}\left(\kappa \frac{\partial T}{\partial z}\right) + \tfrac{1}{2} \mathfrak{F} \epsilon a N(z)\, e^{-\tau/\mu} - R(z)\,. \tag{8}$$

We write n for the number of degrees of freedom, $N(z)$ for the density, κ for the thermal conductivity, $\mathfrak{F}$ for the solar flux (integrated over the relevant spectral range—the factor $\frac{1}{2}$ allows for the heat flow on the nighttime side of the planet), ϵ for the mean efficiency of heating (the fraction of the absorbed solar flux that appears as heat), a for the mean cross-section for absorption (averaged over the spectral range), τ for the optical depth (also averaged over the spectral region), and $\mu = \cos\theta$ for an effective zenith

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angle, θ, of the sun. Here $R(z)$ is the function giving the rate of thermal emission, as in equation (2).

Integrating for a steady state, with $d\tau = N\alpha dz$, gives

$$q(z) \equiv \kappa \frac{dT}{dz} = \tfrac{1}{2}\mathfrak{F}\epsilon\mu(1 - e^{-\tau/\mu}) - \int_z^\infty R(z)\,dz, \tag{9}$$

where q is the heat flux downward. Equation (9) states that the rate of heat flow down the temperature gradient at height z is equivalent to the total energy deposited above z less the amount that is re-radiated per unit time. Numerically, we may write $\kappa = 36 \times 10^{-5}\, T^{3/4}$ erg km/cm^2 sec ° K for N_2; it is about 50 per cent larger for O (see the summary statements by Bates 1959).

In Section III, *a*, a value of $q = 1$ erg/cm^2 sec was adopted for the region just above the earth's mesopause, and this would be equivalent to $\frac{1}{2}\mathfrak{F}\epsilon\mu$, if radiative losses in the thermosphere were small. Radiative cooling is important, of course, near the mesopause but probably diminishes shortly above that level. Owing to the temperature increase, some additional radiative loss may occur by the [O I] emission at 62 μ in the lower thermosphere.

At any rate, the radiative losses in the upper thermosphere are negligible, and there the total conductive flux must be $q = \frac{1}{2}\mathfrak{F}\epsilon\mu \approx 0.8$ erg/cm^2 sec to maintain a gradient of about 20° K/km at 120 km (Bates 1959; Hunt and Van Zandt 1961). Whereas an equivalent amount of energy is deposited through dissociation, the fact that the temperature gradient at 90 km is only some 10° K/km, in spite of a smaller κ, suggests that radiation loss between 90 and 120 km is appreciable.

The temperature increases only slightly beyond the level where τ is a few tenths. Bates (1956) first pointed out from considerations of this nature that if photoionization is the source of thermospheric heating, the temperature must start to level out within the F_1 region and approach a value well below 1500° K. The terrestrial thermosphere shown in Figures 1 and 2 is Bates's (1959) model computed for solar heating and fitted to satellite determinations of mass density.

b) Mars

Temperatures and densities.—On Mars the radiative loss of energy absorbed through dissociation is more severe than on earth. The thermosphere is produced entirely by energy absorbed in photoionization. It begins somewhat above the level of vibrational relaxation and, with the adopted mesosphere model, starts at a very low temperature of 76° K. If there were no radiative loss in the thermosphere, a total conductive flux of only 0.2 erg/cm^2 sec would give an initial gradient of nearly 20° K/km and would raise the temperature to an asymptotic value exceeding 2000° K.

This would probably be the situation if Mars had the same chemical composition as the earth, a trace of CO_2 and a substantial amount of O_2. At these temperatures O would readily escape from the planet (Spitzer 1949; Urey 1959). However, Mars's actual large abundance of CO_2, which becomes CO and O in the thermosphere, provides an effective thermostat. The CO radiates according to equation (2) (Bates 1951), which cools the atmosphere and allows O to be retained (see below). In equation (2) we now write [in place of $f(CO_2)$] $f(CO) = 2 \times 10^{-2}$ and provisionally adopt $\eta = 10^{-14}$ cm^3/sec. The CO vibrational quantum is 0.27 ev.

In the thermosphere the temperature rise is so rapid that the isothermal approximation, as used in equation (3), is quite inaccurate. For a linear increase in scale height according to $dH = \beta\, dz$ for constant β over the interval z_0 to z_1, we obtain

$$\int_{z_0}^{z_1} R\,dz = \tfrac{1}{2} H N^2(z_0)\, f(\mathrm{CO})\, \eta\, h\nu\, e^{-h\nu/kT_0} G(z_0, z_1), \tag{10}$$

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where

$$G(z_0, z_1) = \frac{2}{\beta} \frac{e^{h\nu/kT_0}}{(h\nu/kT_0)^{2/\beta+1}} \int_{h\nu/kT_1}^{h\nu/kT_0} u^{2/\beta} e^{-u} du . \tag{11}$$

The integral may be expressed in terms of incomplete Γ functions and evaluated with Pearson's (1934) tables. In this fashion, one may evaluate changes in the radiation term in equation (9) over a finite interval in height.

A self-consistent model of the thermosphere requires several numerical integrations of equation (9), with the density computed at each step from a well-known formula for constant β (e.g., Nicolet 1960). In successive models one changes the assumed conductive flux at the mesopause until, by trial and error, the final model provides the total amount of radiative loss that was postulated.[2] For example, the model shown in Figures 1 and 2 is for a mesopause flux of 0.07 erg/cm² sec; it yields a total radiative loss of 0.13 erg/cm² sec, occurring mainly in the 150–250-km region, and gives the desired flux of 0.20 erg/cm² sec at high altitude. Still higher, the absorption of sunlight occurs, and the conductive flux is again reduced. The model approaches an asymptotic temperature of about 1100° K, provided that the mean cross-section for absorption of the ionizing frequencies is as low as 3×10^{-18} cm². It is not a straightforward matter to decide how a change in this or that parameter or in the "boundary conditions" at the mesopause will affect the thermospheric model. By carrying through the calculations for a range of basic parameters, I hope to gain some idea of the uncertainties to be ascribed to a model, but this extensive program must be deferred to a later publication.

Oxygen distribution and escape.—Spitzer's (1949) well-known treatment of the escape of planetary gases has been extended by Bates and McDowell (1957, 1959) and Urey (1959) to allow for the sluggishness of diffusion in replacing atoms as they escape from the *critical level*, defined as the height where collisions no longer seriously inhibit escape. Bates and McDowell also introduced other refinements that are useful in calculations of hydrogen and helium escape; it is then helpful, for example, to have an analytical relation between the abundance at the ground and that at the critical level.

The escape flux at the critical level, z_c, where the O density is $N_c(\mathrm{O})$, is

$$F = N_c(\mathrm{O}) \left(\frac{kT_c}{2\pi M}\right)^{1/2} (1 + X)\, e^{-X}, \tag{12}$$

with

$$X = \frac{G\mathfrak{M}M}{kT_c r_c} . \tag{13}$$

Here $\mathfrak{M}$ is the mass of the planet, M the mass of the atom, and r_c the distance of the critical level from the center of the planet. Above the critical level there is, by definition, only one mean-free path or an integrated density of $NH \approx 3 \times 10^{14}$ particles/cm². For Mars we find $z_c = 1500$ km, $T_c = 1100°$ K, $H(\mathrm{O}) = 300$ km, and $N(\mathrm{O}) = 10^7/\mathrm{cm}^3$. (The O density surpasses the N_2 density at 700 km, where there are $2 \times 10^8/\mathrm{cm}^3$ of each species.)

The escape flux at z_c is 3×10^6 atoms/cm² sec; the escape rate above 1 cm² of planetary *surface* is twice this amount. The lifetime for depletion of oxygen to $1/e$ its original abundance is then 10^9 years. Earlier Spitzer (1949) and Urey (1959) noted that a temperature as high as 1500° K seemed inconsistent with the presence of oxygen, even in the form of CO_2. Moreover, it seems likely that the addition of substantially more oxygen

[2] In fact, each step of the integration proceeds by trial and error, since one must assume an average value of β over the next interval in order to evaluate the radiation term (10). Then when dT/dz is obtained from equation (9), the assumed β may turn out to be in error, necessitating a revision. Incidentally, the relation between H and T (and between β and dT/dz) must allow for the important variation of gravitational acceleration, g, with height.

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to the atmosphere would raise the altitude at which solar energy is deposited and would thereby raise the temperature of the exosphere. This suggests that Mars could never have retained large quantities of O_2 over long geological periods, since any such oxygen would have induced its own rapid escape.

Ionosphere.—The terrestrial E region appears to be essentially a Chapman (1931) layer, with electrons disappearing through dissociative recombination. Free electrons are produced by X-rays acting on all species and by ultraviolet photoionization of O_2. A Chapman layer has its ionization peak at unit optical thickness, so that portion of the Martian E region arising from X-rays (here called E_2) appears at the same value of NH as it does on earth. Hence the E_2 region maximum is at 200 km (also see Goody 1957).

For O^+ and N_2^+ recombination will thus occur by

$$N_2^+ + e \rightarrow N + N \tag{14}$$

or by

$$O^+ + N_2 \rightarrow NO^+ + N\,, \tag{15a}$$

followed by

$$NO^+ + e \rightarrow N + O\,. \tag{15b}$$

For recombination of O, the N_2 densities are sufficiently high that formula (15*a*) will proceed quickly and (15*b*) will be the limiting process. The electron production rate is less on Mars than on earth because of both the smaller flux and smaller atmospheric density at the height of the E_2 region. Assuming that the recombination coefficients are about the same in the two atmospheres, we obtain electron densities less than on earth by a factor of 3. Thus the E region has electron densities at noon of less than 5×10^4/cm^3; just how much less depends on how much of the terrestrial E region is produced by X-rays.

On Mars photoionization of O_2 ($h\nu > 12.0$ ev) will occur near the mesopause, where O_2 is formed. This region we shall call E_1; it occurs at O_2 densities of the order of 10^{12}/cm^3, slightly larger than at the E peak on earth, which tends to compensate for the reduced flux. However, at the low mesopause temperature, dissociative recombination is probably rapid, and it is unlikely that N_e could exceed the noon terrestrial value of 10^5 electrons/cm^3. Nevertheless, the E_1 region is likely to be quite important on Mars: occurring at total densities of 10^{14} particles/cm^3, it may be expected to absorb radio frequencies in the same manner as the terrestrial D region.

The terrestrial F_1 region is also like a Chapman layer. Its basic mechanisms are thought to be similar to those in the E region. Production of O^+ by ultraviolet light ($h\nu > 13.5$ ev) is followed by recombination processes (15*a*, *b*). In addition, ionization of N_2 ($h\nu > 15.5$ ev) produces additional electrons, which disappear readily by dissociative recombination (14).

On Mars an F_1 region analogous to the one on earth is expected around 320 km, with a noon density of about 10^5 electrons/cm^3. Like the E regions, it disappears soon after sunset.

The F_2 region on earth is thought to arise from the same ionization processes that produce F_1. The distinction is merely that, at high altitudes, reaction (15*a*) proceeds very slowly and becomes the limiting process in recombination. Ion production occurs at very small optical thicknesses, and little solar heat is put into the atmosphere in F_2. But recombination is so slow that the electron density is high and the ionization largely persists through the night. Another important feature of F_2, however, is that local ionization equilibrium is not strictly followed. Instead, diffusion of ions plays an important role in determining the profile of electron density.

With local ionization equilibrium, the condition for (15*b*) to remove electrons as fast as NO^+ ions are formed by (15*a*)—and hence the transition from F_1 to F_2—would occur on Mars around 700 km. Here the electron density would be some 4×10^4/cm^3 and

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would increase gradually with greater height. The total particle density there (about $4 \times 10^8/cm^3$) and above is so low, however, that the ionic density distribution will be largely governed by diffusion. It appears, therefore, that an F_2 region will not develop to the extent that it does on earth and will decay more rapidly at night.

The research described in this paper is being supported in part by grant NsG 118-61 from the National Aeronautics and Space Administration. Mr. J. L'Ecuyer kindly assisted with the computations.

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