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THE ATMOSPHERE OF MERCURY

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1. Introduction

Mercury is the smallest planet of the solar system and is also nearest to the sun. The combination of these two circumstances makes it a hot and gravitationally "weak" planet. It has therefore been a firm belief that any atmosphere which the planet could have acquired during its formation and/or by the subsequent outgassing processes would be difficult to hold. High thermal velocities of the particles in the atmosphere would win over the relatively small escape velocity and therefore, shortly after its formation, the planet would lose practically all its atmosphere to the space.

In the literature on planets, the description of the atmosphere of Mercury has usually been very brief (e.g., KUIPER, 1952; UREY, 1959; DOLLFUS, 1961), and Mercury has invariably been compared with the moon, as a planet without an atmosphere. In the last few years, however, evidence has accumulated that the atmosphere of Mercury may not be, after all, as tenuous as that of the moon, but could have a surface pressure of as much as 1–10 mb, at least 10^{10} times greater than at the surface of the moon. This has been further substantiated by the recent identification of a small amount of CO_2 in the atmosphere of Mercury (MOROZ, 1963, 1965).

Also, the recent radar investigations of the planet (PETTENGILL and DYCE, 1965) have led to the remarkable discovery that Mercury is not in synchronous rotation but, instead, rotates much faster, with a period of 59 ± 5 days.

It is the purpose of this paper to review the available information on the temperature, composition, and surface pressure of Mercury's atmosphere, and to discuss the implications of a non-synchronous rotation of the planet on the atmosphere. We construct several models for the atmosphere which are consistent with the observations in an attempt to determine if they are stable against depletion of the atmosphere by gravitational escape. The results are discussed from the standpoint of the origin of the atmosphere.

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2. Rotation

2.1. Early visual observations of the planet

The first telescopic observations of Mercury, leading to the discovery of the phases of the planet, were carried out by Zupus in 1639. Because Mercury is small, distant, and is never more than 28° of arc away from the sun, it is extremely difficult to observe, being only visible low on the horizon in the very early hours of twilight. It is for this reason that it was not until 1882 that the first reliable map of the planetary disk was drawn. Schiaparelli had then conceived the idea of observing the planet in the daytime. But because of this inherent difficulty in observing the planet, the data on Mercury, even to this day, remains extremely meager.

The drawing of Mercury made by Schiaparelli (1889) showed irregular dark markings on the surface of the planet which, after subsequent observations, were found to be of a permanent nature. Antoniadi (1934), after an extensive series of observations, produced the first planisphere of the day-side of the planet. Lyot (1943) and Dollfus (1953) continued such observations of Mercury and produced a series of drawings which have been discussed in detail by Dollfus (1961).

From the first observations, it was noticed that when the planet was continually observed over a period of several hours, the markings did not shift across the disk. This led Schiaparelli to conclude that Mercury must have a synchronous rotation, spinning on its axis in 88 days which is also the time of its orbital revolution. When later observations revealed a recurrence of the markings, even over long intervals of time, it led practically all observers to conclude that Mercury had a rotation synchronous with its revolution. This conclusion was further substantiated by the theoretical argument that because of the nearness of the planet to the sun, the solar tidal forces would be so strong that they would slow down the rotation of Mercury in the very early history of the solar system. Like the moon with respect to the earth, Mercury would be "locked" in the direction of the sun.

2.2. Recent radar measurements

During the inferior conjunction of Mercury in April, 1965, Pettengill and Dyce (1965) made radar observations of the planet and derived a value for the rotation period of 59 ± 5 days. Figure 1 shows their results of the limb-to-limb doppler spread as a function of time for four measurements taken during the months of March and April, 1965. The best fit curve indicates direct rotation with the sidereal period of 59 ± 5 days. The authors have also pointed out that the direction of the pole may be normal to the planetary orbit.

This new result is in complete disagreement with the hitherto widely accepted value of 88 days, derived from the visual observations as discussed in the previous section. McGovern, Gross and Rasool (1965) have investigated the reasons for this apparent inconsistency between the results of the early visual observations and the recent radar measurements. Nearly 50 drawings of Mercury published by Lowell (1902), Antoniadi (1934), Lyot (1943), Dollfus (1961) and Baum (see Sandner,

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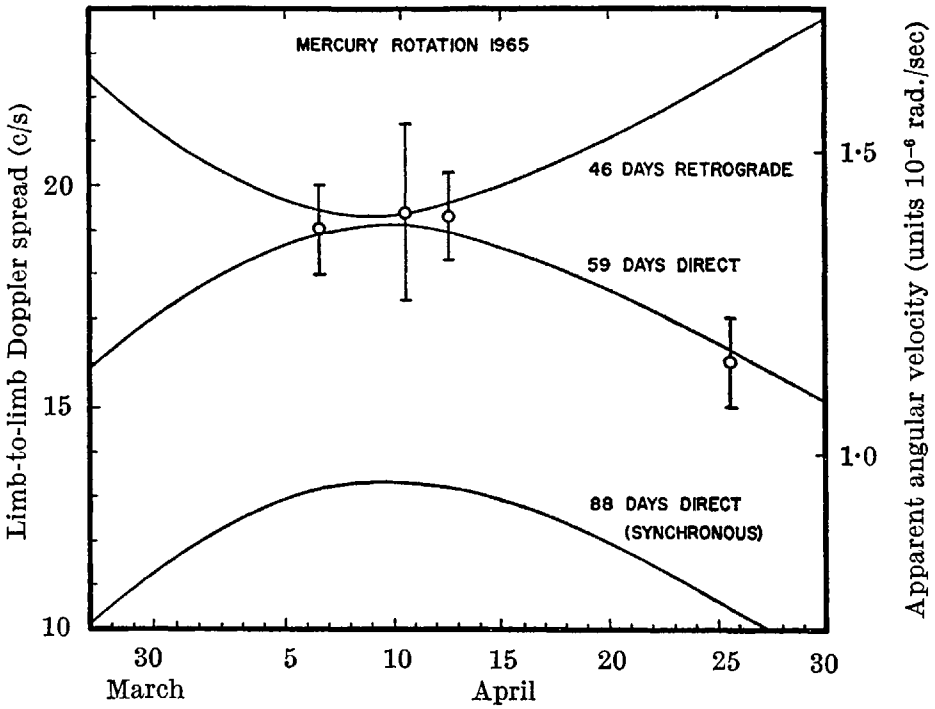

Fig. 1. Plot of the apparent rotational angular velocity of the planet Mercury versus date for several values of rotation during the inferior conjunction of April 1965. The values inferred from the measurements are shown with their estimated errors (after PETTENGILL and DYCE, 1965).

1963 and MOORE, 1960) were examined. Of these six pairs of drawings showed near duplication of markings and phase. These pairs are listed in Table I.

TABLE I

Pairs of similar drawings of Mercury

Reference	Time interval T (in days)
1. ANTONIADI (Aug. 11, 1924) and ANTONIADI (June 21, 1927)	1044
2. ANTONIADI (Aug. 23, 1927) and ANTONIADI (Aug. 6, 1928)	349
3. ANTONIADI (Oct. 4, 1927) and ANTONIADI (Aug. 23, 1929)	690
4. ANTONIADI (Aug. 6, 1928) and ANTONIADI (July 20, 1929)	348
5. LYOT (July 22, 1942) and DOLLFUS (Oct. 12, 1950)	3004
6. BAUM (March 15, 1952) and BAUM (March 1, 1953)	351

The authors pointed out that duplication of markings and phase for any single pair of drawings of Mercury does not necessarily indicate that synchronous rotation of the planet is the only possible solution. A number of other periods of rotation are possible as given by the following equation:

$$P = \frac{T}{n + \theta/360} \tag{1}$$

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where P is the period of rotation of Mercury in days, T the time interval between the two observations made at the same phase, n the number of complete rotations of the planet during this time interval and θ the average angular displacement of the earth and Mercury in their orbits. Figure 2 shows all possible values of P between 40 and 90 days as calculated for each pair of drawings. It is noted that in addition to an 88-day period, there are at least four more values of P, viz., 43.6, 50.1, 58.4

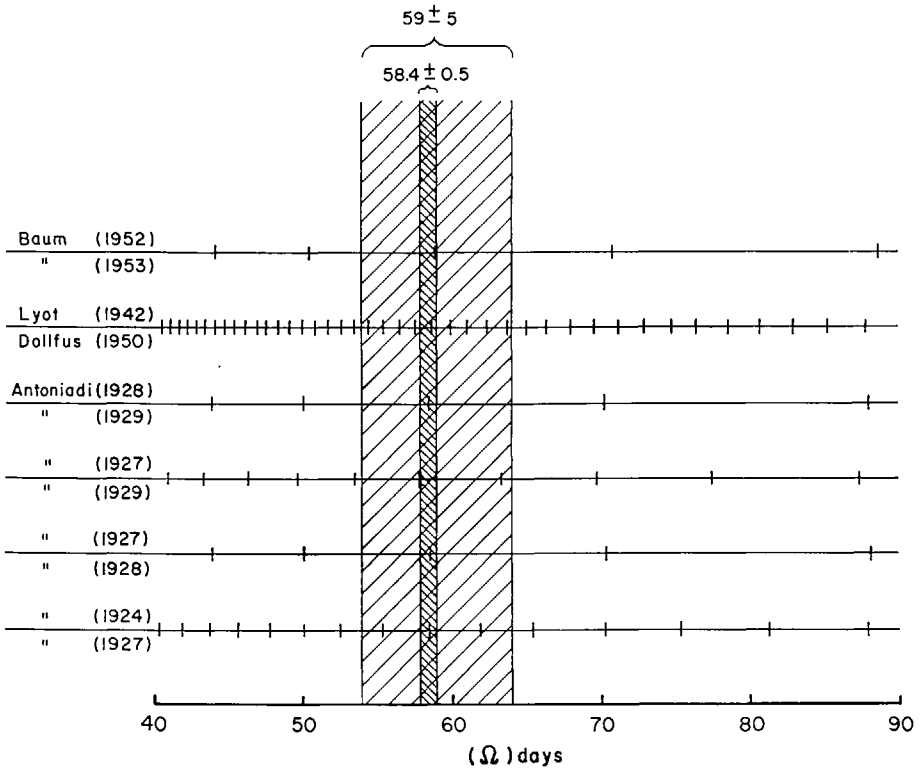

Fig. 2. Rotation periods of the planet Mercury, in days, as derived from six pairs of drawings. The single-hatched area shows the limits in the rotation period allowed by the radar observations of Mercury. Visual observations indicate a value of 58.4 ± 0.5 (double-hatched area).

and 70.2 days which will be consistent with all the six pairs of drawings. However, only one of these values is within the allowed limits of 59 ± 5 days determined by radar. Therefore the rotation period of 58.4 ± 0.5 days is consistent with both the visual and the radar observations of Mercury.

If, however, we take only the pair number 5, which has the longest time interval and also the most convincing duplication of markings, we can narrow down the accuracy limits to a value of 58.65 ± 0.10 days.

PEALE and GOLD (1965) have made an interesting analysis of the problem and have explained the non-synchronous rotation of Mercury by the following simple argument. The ellipticity of the orbit of Mercury around the sun is very large, about 0.2. The tidal torque is proportional to $1/r^6$ and will be greatest at the perhelion. The planet can therefore acquire a rotation period lying between 56.6 and 88 days, depending on the dissipation function of the tidal forces. If the value of 58.65 ± 0.1

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days, as derived above, is correct, then following Peale and Gold's argument, it will indicate that there is a significant amplitude dependence of the tidal dissipation on Mercury.

Another argument in favor of the 58.6 day period has been advanced by COLOMBO (1965), who suggests that a period of rotation which is exactly 2/3 of the orbital period will be stable because such a rotation period will bring the axis of minimum moment of inertia in line with the sun-Mercury radius vector at every perhelion. However, based on only this line of argument, one can obtain a large number of "stable" rotation periods for Mercury. Combining such an analysis with the arguments given by Peale and Gold and also with the radar and visual observations, one can finally obtain a precise value for the rotation, viz. 58.65 days.

The non-synchronous rotation of Mercury implies, contrary to the previous belief, that all parts of the planet are exposed to the sun at one time or another. The length of the day on Mercury is approximately 176 earth days, and it may be a fair assumption that the surface temperature on the night side of the planet may, after all, not be as low as 28 °K which was calculated by WALKER (1961) for a synchronously rotating planet.

3. Evidence for the Presence of an Atmosphere on Mercury

The astronomical constants of the planet Mercury are summarized in Table II.

TABLE II
Astronomical constants of Mercury

Mass	0.054 (Earth 1)
Radius	0.39 (Earth 1)
Density	5.05 gm/cm^3
$V_c = \sqrt{\frac{2MG}{R}}$	4.2 km/sec (Earth 11.3 km/sec)
g	350 cm/sec^2
Distance from sun	0.39 AU
Solar radiation flux	6.6 × at Earth
Albedo	0.06 (Earth 0.4)
T_e (sub-solar point)	616 °K
Orbital period	88^d
Rotation	58.65 ± 0.1^d

3.1. POLARIMETRIC STUDIES

One of the most important techniques of studying the surface of another planet is the observation of the polarization of solar radiation reflected by the planet. This technique was developed by LYOT (1929) and has since been used extensively by him and later by Dollfus to study the surfaces of the moon and planets. A detailed discussion of this method is given by DOLLFUS (1957, 1961, and 1962).

One of the by-products of such an observation is an estimate of the total atmosphere of the planet. By comparing the polarization of the solar radiation as reflected

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from the center of the planetary disk and from the limb, one can obtain an estimate of the scattering of the radiation by the atmosphere. Making such measurements at different wavelengths and assuming Rayleigh scattering to be the main process, one can estimate the total number of molecules in the atmosphere. In this manner, LYOT (1929) gave an upper limit on the extent of the atmosphere of Mercury, viz., $<21/1000$ of the earth's atmosphere. For the smaller value of g for Mercury, this would correspond to a pressure of 7 mb. In these calculations, it was assumed that the albedo of Mercury, in the visible, was 0.16. The recent measurements by HARRIS (1961) give a value of 0.063. If we correct Lyot's calculations for this new value of albedo, we obtain an upper limit on the surface pressure on Mercury of 3 mb. Dollfus repeated Lyot's observations with an improved polarimeter and using the same method of reduction of data, gave a value of the atmospheric pressure at the surface of Mercury equal to 1 mb within an order of magnitude, i.e., $P_S = 10^{0 \pm 1}$.

It is difficult to accept the results without qualifications because of the inherent limitations of this method in deducing the atmospheric pressure. The most important source of error is the assumption that the scattering is entirely molecular. Any particulate matter which may be present in the atmosphere will result in a considerable over-estimate of the pressure. Also, it is assumed that the atmosphere of Mercury has the same scattering properties as air. If, however, the major component of the atmosphere of Mercury were not nitrogen but, for example, neon, the pressure quoted above would be an under-estimate by as much as a factor of 10. CHAMBERLAIN and HUNTEN (1965) have recently given a critical review of the inherent difficulties in this method for estimating the atmospheric pressure. Accepting their arguments, we would conclude that the uncertainty of an order of magnitude quoted by Dollfus may be quite optimistic, and the actual atmospheric pressure on the surface of Mercury may be anywhere between 0.01 and 10 mb.

3.2. SPECTROSCOPIC STUDIES

The search for atmospheric gases on Mercury was carried out by ADAMS and DUNHAM (1932) and later by KUIPER (1952), but failed to give any positive results. It was therefore generally believed that if an atmosphere exists on Mercury, it must be composed of rare gases, perhaps argon (FIELD, 1964), not observable by spectroscopic techniques. Recently, however, MOROZ (1963) has detected the presence of CO_2 in the atmosphere of Mercury. He recorded the spectrum of solar radiation as reflected by Mercury and compared it with that of the moon. The equivalent width of the 1.6 μ CO_2 band appears enhanced in the Mercury spectrum, indicating the presence of CO_2 in the atmosphere of the planet. This is illustrated in Figure 3. A new series of similar measurements, however, show a lesser enhancement of the equivalent width (MOROZ, 1965).

As the strengths of the saturated CO_2 lines in the 1.6 μ band depend both on the effective pressure and the amount of CO_2, it is not possible to separate the two by observing only in this one band. SPINRAD, FIELD and HODGE (1965), therefore, attempted to measure the intensity of the weak unsaturated lines of CO_2 at 8700 Å

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which are not pressure-dependent. Combining these measurements with those of Moroz, they expected to get the amount of CO_2 and the total atmospheric pressure at the surface of the planet. Despite a detection limit of as low as 4 mÅ, they were unable to observe CO_2 lines in the 8700Å region. They therefore proposed an upper limit to the CO_2 content of 57 m-atm., which corresponds to a maximum possible

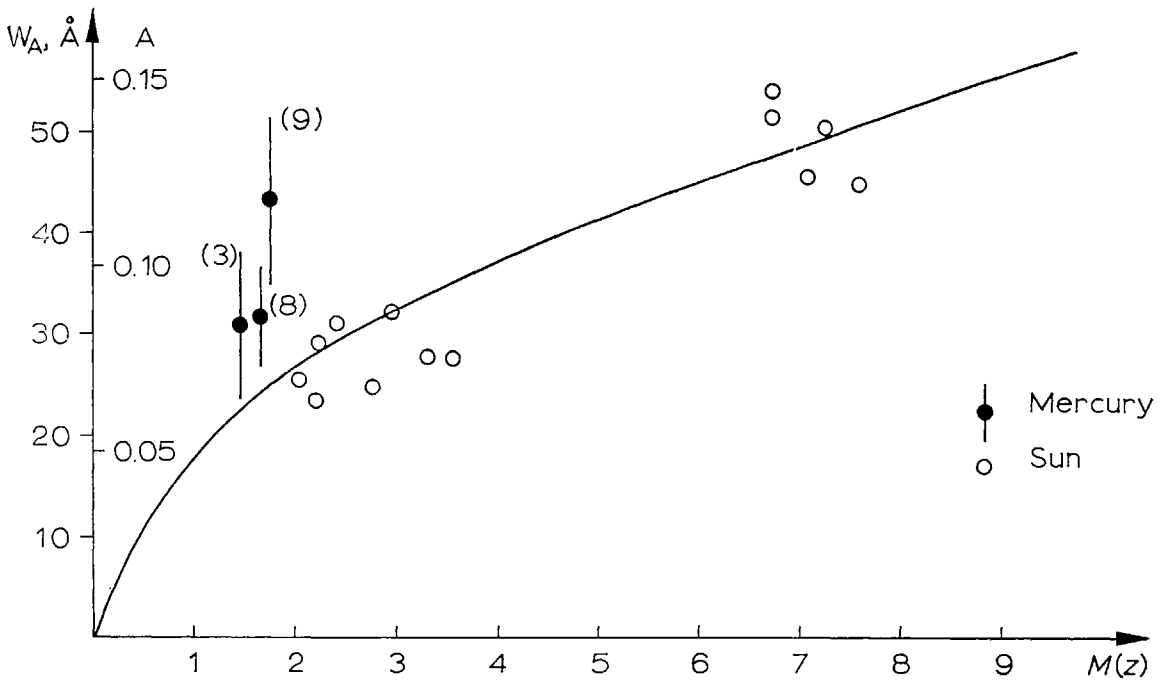

Fig. 3. The total equivalent width of CO_2 bands at 1.59 μ in the spectrum of Mercury and the sun. The figure in parentheses indicates the number of observations on which the value is based. The solid curve is the theoretical curve of growth for CO_2 absorption in the earth's atmosphere (after Moroz, 1963).

partial pressure of 4.2 mb on the surface. Combining this upper limit on the amount of CO_2 with Moroz' observation of the equivalent width of the 1.6 μ CO_2 band, Spinrad, Field and Hodge conclude that the surface pressure on Mercury is near 4 mb if the atmosphere is pure CO_2, and higher if other gases are present.

Kozyrev (1964) observed Mercury during an eclipse of the sun and found emission lines of hydrogen in the spectrum. He therefore postulated that the atmosphere of Mercury may contain a significant amount of hydrogen. Spinrad and Hodge (1965) have given a more acceptable explanation of these line profiles, suggesting that they could have been formed by a spurious blending of the doppler-shifted Fraunhofer lines in the spectrum of Mercury and the unshifted component in the sky spectrum.

In summary, it appears from the evidence described above that Mercury has a tenuous atmosphere, with a surface pressure between 1 and 10 mb and containing a small amount of CO_2.

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4. Temperature Measurements

For a slowly rotating planet like Mercury, the effective equilibrium temperature on the day side of the planet is given by

$$T_e \sim \left(\frac{F}{2\sigma}(1-A)\right)^{1/4} \tag{2}$$

where F is the solar flux reaching Mercury (9.2×10^6 ergs/cm^2/sec) and A is the albedo (0.063) (HARRIS, 1961). For these values of F and A, the mean effective temperature of the day side of Mercury is 520 °K. At the sub-solar point the temperature will be 616 °K. A tenuous atmosphere precludes any substantial greenhouse effect and therefore we can expect the surface temperature of the day side to be approximately the same as the effective temperatures computed above.

Temperature measurements of Mercury have been made by recording the thermal emission of the planet in the infrared and microwave regions of the spectrum. We shall discuss the results of these two types of observations in the following paragraphs.

4.1. INFRARED

PETTIT (1961) has measured thermal emission from Mercury in the 8–12 μ region and obtains a sub-solar point surface temperature of 615 °K, in close agreement with the effective temperature given above. This agreement between the observed surface and effective temperatures of the planet indicates that the surface of Mercury radiates as a black body in the far infrared with an emissivity of approximately unity.

4.2. MICROWAVES

Thermal radiation from Mercury in the microwave region at $\lambda=3$ cm was measured by HOWARD, BARRETT and HADDOCK (1962). Figure 4 shows their temperature values as a function of phase angle. The interesting feature of their results is that the dependence of the temperature on the phase angle is not so strong as would be expected if the dark side were extremely cold, at a temperature close to 28 °K (WALKER, 1961).

FIELD (1964) interpreted these results to suggest that the dark side of Mercury may actually be at a higher temperature. Curve A in Figure 4 shows the variation of temperature as a function of phase angle which would be observed if the dark side were at a temperature as high as 300 °K. The observed values of temperature show a better agreement with curve A than with curve B which is computed for a dark side at 0 °K. Field also suggested that in order to explain these observations, an atmosphere must be present which would transport the heat required to raise the temperature of the dark side to 300 °K. This suggestion of Field's is based on the assumption that the measured radio brightness temperatures are actually the surface temperatures of Mercury.

MAYER (1961) has however pointed out that the thermal emission in the radio

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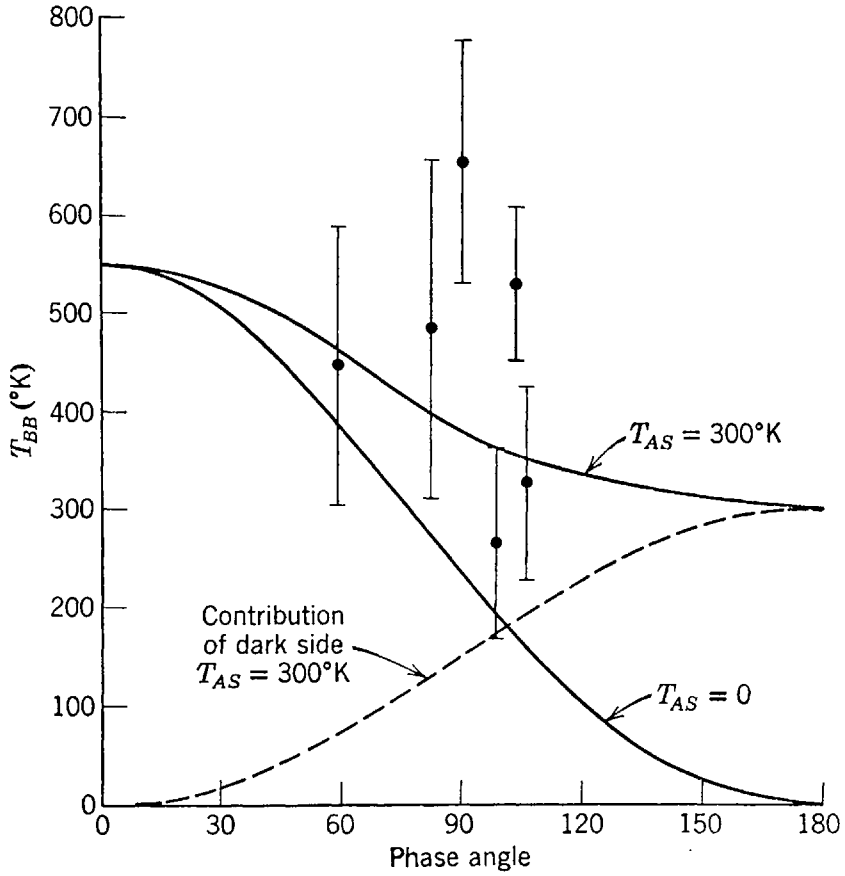

Fig. 4. Computed apparent blackbody temperatures of Mercury at centimeter wavelengths. Lower solid curve (B) applies if dark side is cold. Upper solid curve (A) includes contribution of dark side at 300 °K (after FIELD, 1964).

wavelengths originates a few wavelengths below the surface of the planet. More recent radio measurements of Mercury by KELLERMAN (1965) at $\lambda = 11$ cm corroborate this argument. Figure 5, taken from Kellerman, does not show any phase effect at all, indicating that the temperature a few decimeters below the surface of Mercury may be constant at 300 °K, both on the day and night sides. Now that it is known that the rotation of Mercury is non-synchronous, and that all parts of the planet are exposed to solar radiation every 88 days (one solar day on Mercury is equal to 176 earth days), a 300 °K temperature in the sub-surface layers on the dark side of Mercury seems plausible.

No observational information is yet available on the actual surface temperature on the dark side of the planet, but Field has argued that if argon is present in the atmosphere, then the temperature of the surface of the dark side should be higher than 56 °K so that the atmosphere does not freeze out. For a carbon dioxide atmosphere, this temperature should be higher than 150° K.

MINTZ (1962) has made a theoretical study of the problem of heat transport across the terminator by winds, for a slowly rotating planet. Using his expression which relates the wind velocity at the terminator and the temperature on the dark

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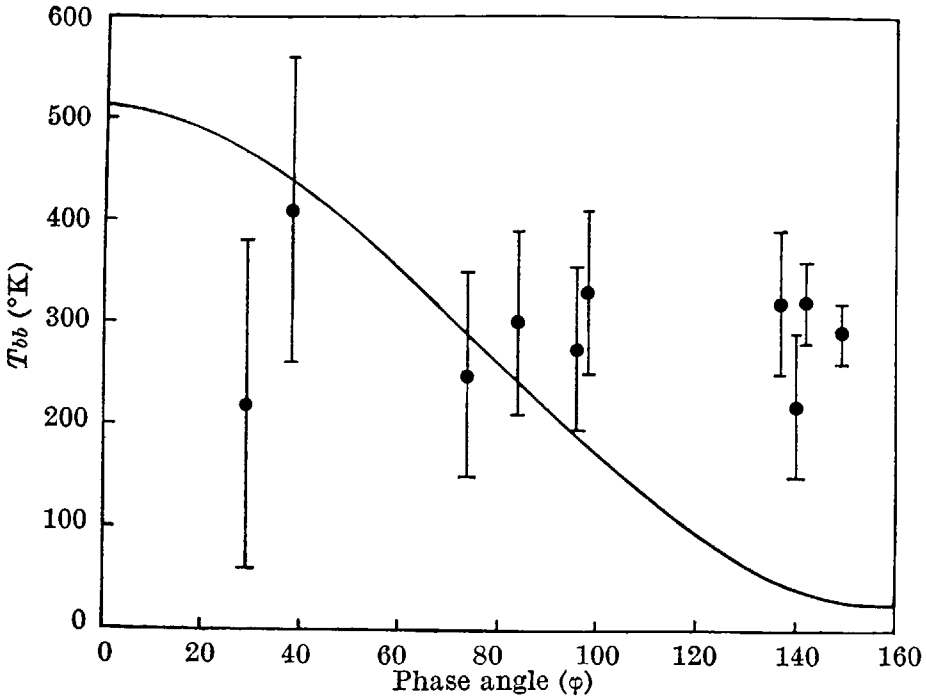

Fig. 5. Equivalent blackbody disk temperature of Mercury as a function of planetocentric phase angle. The solid line shows the expected phase law for a smooth non-rotating planet with no atmosphere and a low thermal conductivity. The curve shown corresponds to a sub-solar temperature of 610 °K and it is assumed that there is no source of heat other than solar radiation (after Kellerman, 1965).

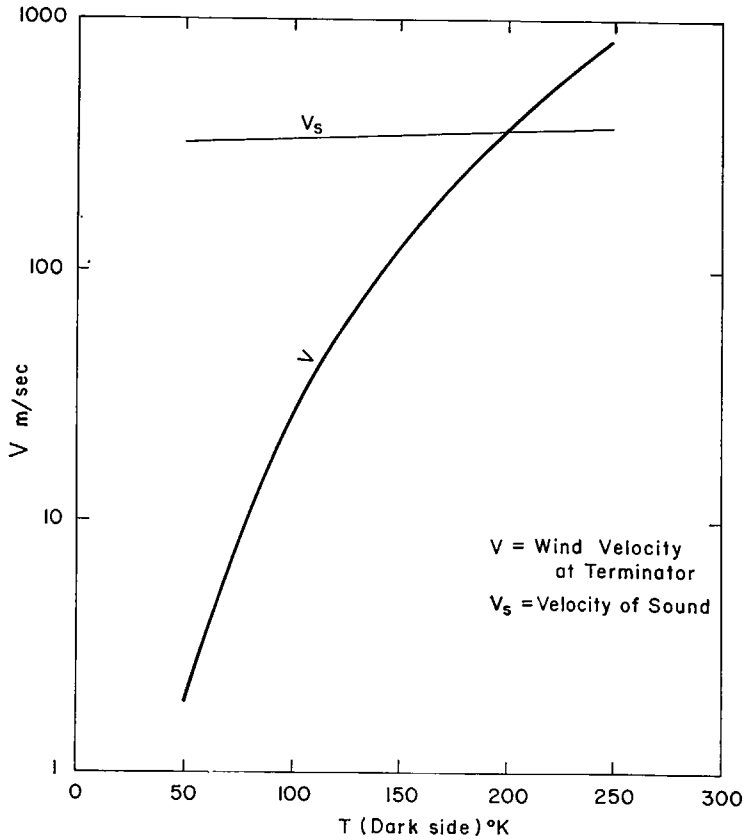

Fig. 6. The required wind velocities as a function of temperature increase on the dark side of the planet Mercury. V_s = velocity of sound in the atmosphere at these temperatures.

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side of the planet, we have made a sample calculation for the Mercurian atmosphere, mainly composed of CO_2, and for $P_S = 5$ mb. Figure 6 is a plot of the required wind velocities as a function of temperature increase on the dark side of the planet. Also plotted is a curve giving the velocity of sound in the atmosphere at these temperatures. It appears that even in a thin atmosphere of 5 mb surface pressure, atmospheric circulation is an efficient means of transporting heat from the day side to the night side of the planet. A 120 m/sec wind can maintain the dark side of Mercury at 150 °K, thus preventing CO_2 from freezing out. Wind velocities of the order of 100 m/sec are frequently observed in the upper atmosphere of the earth.

5. Problem of Gravitational Escape and Accumulation of an Atmosphere

In this section we shall examine the processes which govern the evolution of an atmosphere on a small and hot planet like Mercury. There are two schools of thought on the problem of the origin of a planetary atmosphere.

It has been suggested that the present atmosphere is either (a) the remnant of a dense gaseous envelope which the planet may have acquired at the time of its accumulation out of the contracting solar nebula, or (b) the result of slow outgassing from the interior of the planet via, e.g., volcanic activity and the exhalation of gases through fissures in the crust.

For the present atmosphere to be the remnant of a primitive one, it must be stable against the two important loss processes: thermal escape of gases from the gravitational field of the planet, and the dissipative effect of the solar wind. The atmosphere, in this case, would be composed of those gases which have not yet escaped or been ejected by the solar wind. Their abundances would, therefore, still be in the same relative amounts as observed in the sun.

On the other hand, it is difficult to ascertain if the present atmosphere of another planet is the result of outgassing. Even in the case of the earth, the gas exhalation rates from the crust are very poorly known and extrapolating these to other planets could be very unreliable. The volcanic outgassing not only depends on the differential processes and tectonic activity occurring in the interior of the planet, but also on the nature of "volatiles" which were trapped within the planet at the time of its formation. These are difficult questions in cosmogony and are beyond the scope of this paper.

We would like to point out that the major argument in favor of the outgassing hypothesis is the demonstration that the present atmosphere could not be primordial in origin.

For this purpose we shall discuss the problems of gravitational escape and the effect of solar wind in more detail, with particular reference to Mercury.

5.1. Thermal escape

The flux of particles escaping from the top of a planetary atmosphere is readily estimated from the theory of Jeans (1916) as modified by Spitzer (1952). Neglecting

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diffusion, the total number of escaping particles, F, is given by

$$F = \frac{4\pi R_0^2 n_{0i} C Y e^{-Y}}{\sqrt{6\pi}} \left[\frac{R_c}{R_0}\left(1 + \frac{R_c}{Y R_0}\right)\right] \text{cm}^{-2}\ \text{sec}^{-1} \tag{3}$$

where R_0 is the radius of the planet, R_c is the planetary radius at the base of the exosphere, n_{0i} is the density which would exist at the surface of the planet if the atmosphere were isothermal at all heights, C is the root-mean-square velocity of the atmospheric particles, and

$$Y = \frac{GmM}{kT_c R_0}, \tag{4}$$

where m is the mass of the escaping constituent, M is the mass of the planet, G is the gravitational constant, and T_c is the temperature of the exosphere.

The number of escaping particles therefore particularly depends on the temperature at the top of the atmosphere, on the dimensions of the planet, and also on the molecular weight of the escaping particles. Using Equation (3) we can calculate, for different values of T_c, the time in which the abundance of a gaseous constituent on Mercury would decrease to $1/e$ of its initial value, which is essentially the half-life of the constituent. The results are plotted in Figure 7. The times of escape have been calculated for three different temperature values, viz., 500 °K, 1000 °K, and 2000 °K, and are plotted as a function of atomic or molecular weight of the escaping particles. For a planetary lifetime of 5×10^9 years, it is seen that Mercury will lose all its hydrogen and helium at an exospheric temperature > 200 °K, but will retain atomic

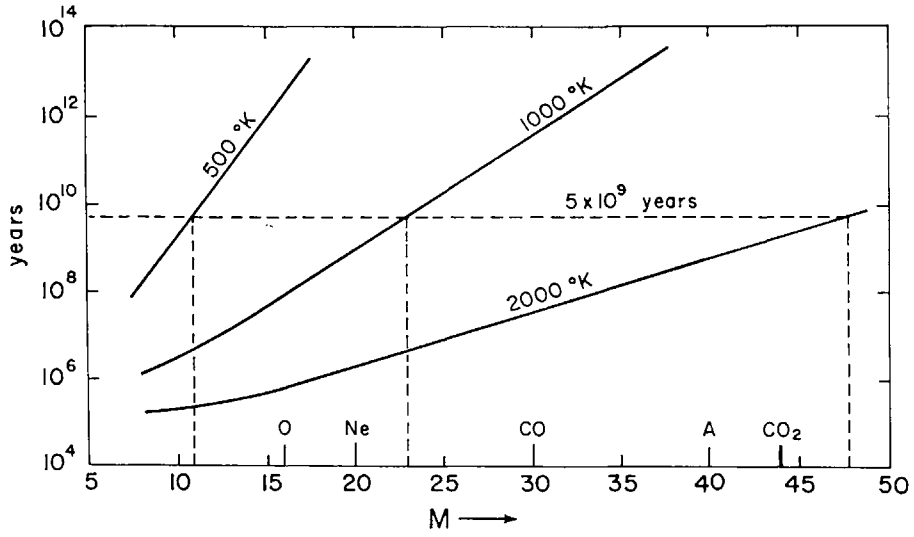

Fig. 7. The time of escape of gases from the planet Mercury as a function of atomic or molecular weight of the escaping particles.

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oxygen at $T_c < 800$ °K, and argon at $T_c < 1800$ °K. At temperatures higher than 2000 °K, only heavy gases like krypton and xenon could be retained by the planet.

Because Mercury is very close to the sun ($R \approx 0.4$ AU) and therefore receives 6.6 times more solar flux than the earth (Table II), one would expect the exospheric temperature to be much higher than for the earth (1500 °K). In fact, the values for T_c for Mercury quoted in the literature are as high as 5000 °K (UREY, 1959). According to Figure 7, at such a high temperature, even xenon will be lost, and one would not expect even a trace of an atmosphere on Mercury. For the atmosphere of Mercury to be stable against gravitational escape, the exospheric temperature should be considerably less than 5000 °K. From the table of solar and cosmic abundances (CAMERON, 1963) the most abundant molecules in a primitive atmosphere, after the escape of hydrogen and helium, would be carbon dioxide, neon and nitrogen. Their relative amounts by volume are given in Table III.

TABLE III

Composition of a primitive atmosphere after the escape of H_2 and He

CO_2	60 %
Ne	25 %
N_2	15 %
A	< 0.5 %

This table has been computed with the assumptions that: (1) all hydrogen and helium have completely escaped, including that hydrogen which could have been in the form of CH_4, NH_3, or H_2O; (2) the oxygen available for the atmosphere has been depleted by that amount which would be required to form metal oxides and silicates; (3) only that much carbon is included in the atmosphere which will combine with the remaining oxygen to form carbon dioxide; and (4) carbon dioxide has not been removed from the atmosphere by reactions with the crust of the planet.

In order for this atmosphere to be stable against gravitational escape, the temperature of the exosphere of Mercury must be <800 °K (Figure 7), which is the escape temperature for atomic oxygen, a dissociation product of carbon dioxide. If, on the other hand, the temperature of the exosphere is higher than 800 °K, then, to maintain a carbon dioxide atmosphere on Mercury, a considerable amount of outgassing will be required to counter the high flux of escaping particles.

5.2. SOLAR WIND

Interactions between the solar wind and the atmosphere of Mercury may produce several important effects such as ejection of ionized particles, intense heating, and ionization in the upper atmosphere, and also accretion of constituents into the atmosphere. But these effects will be highly dependent on the extent and nature of the planetary magnetic field. In the absence of any information on the magnetic field of Mercury, we shall limit our discussion of this aspect of the problem to the

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derivation of a lower limit to the field strength which would protect the atmosphere of Mercury from the dissipative effects of the solar wind.

We shall assume that an effective protection of the atmosphere from the solar wind will be achieved if the magnetic pressure at the height of the exosphere is greater than or equal to the dynamic pressure of the solar wind; that is, we set

$$\frac{B_c^2}{8\pi} \geqslant nmv^2 \tag{5}$$

where B_c is the field strength at the boundary of the magnetosphere, which is assumed to be one planetary radius away from the surface and therefore above the exosphere, and n, m, and v are the number density, mass and velocity, respectively, of the particles in the solar wind. Extrapolating the flux data obtained from Mariner II to the distance of Mercury, we find that $n \sim 100/\text{cm}^3$ and $v \sim 5 \times 10^7$ cm/sec. This gives $B_c \sim 300\ \gamma$ at one planetary radius away from the surface. The dipole field at the surface will then be 0.024 gauss. This is approximately 20 times less than the magnetic field strength of the earth.

The real problem of interaction of the solar wind with a planetary magnetic field is much more complex and is not completely understood. We shall therefore refrain from any comment on the possible mass interchange which may take place between atmosphere and solar wind.

6. Thermal Structure of the Atmosphere

In this section we shall investigate the thermal structure for various models of the atmosphere of Mercury. The main purpose is to determine if the exospheric temperature, in any of these models, is low enough for the atmosphere to be stable against gravitational escape. For these calculations we first derive the temperature of the lower atmosphere of Mercury which is assumed to be in radiative equilibrium. We then discuss the region of the mesopause which separates the lower and the upper atmospheres. The calculation of the structure of the thermosphere and exosphere is then presented.

6.1. Atmospheric models

Four models for the composition of Mercury's atmosphere have been investigated: (a) Pure argon, as suggested by Field (1964), $P_S = 1$ mb; (b) 50% argon, 50% carbon dioxide, $P_S = 5$ mb (atmosphere resulting from outgassing); (c) 60% carbon dioxide, 25% neon, 15% nitrogen, $P_S = 5$ mb (remnant of primitive atmosphere); (d) Pure carbon dioxide, as suggested by Moroz (1965), $P_S = 1$ mb.

6.2. Lower atmosphere and mesopause

The temperature structure of a planetary atmosphere which is in radiative and convective equilibrium has been discussed in detail by Goody (1964). For an optically thin atmosphere the lower region (troposphere) will be in convective equilibrium and

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the temperature gradient close to adiabatic. In the upper region (stratosphere) where the optical thickness in the infrared approaches zero, the atmosphere becomes isothermal at $T_0 = T_e/2^{\frac{1}{4}} = 438$ °K.

The values of the adiabatic lapse rate and the height of the tropopause for model atmospheres (b), (c), and (d) are given in the first two columns of Table IV. The model (a) corresponds to a pure argon atmosphere. Being an inert gas, argon is radiatively inactive in the infrared and therefore the optical thickness of the atmosphere is zero. Hence the temperature structure of the lower atmosphere will not be governed by radiation but by conductive and convective processes. As will be shown later, argon will nevertheless absorb the solar ultraviolet in the upper atmosphere, but because of the radiative inertness, this energy will have to be conducted downwards all the way to the surface to be radiated away from the ground. This will probably tend to establish a positive temperature gradient throughout the atmosphere. For this model, therefore, the temperature profile in the lower atmosphere is highly uncertain, and we shall confine our attention to the exosphere alone, discussed in a later section.

If there are no sources or sinks of energy in the stratosphere, then the temperature for models (b), (c), and (d) will remain constant at T_0 until the level at which the density becomes so low that LTE breaks down, which is denoted as the level of vibrational relaxation. At this level in the atmosphere, the radiative lifetime of CO_2 is shorter than the vibrational relaxation time. Assuming a relaxation time for CO_2 of 3 μsec at 420 °K, the pressure level at which vibrational relaxation will occur is determined to be 8 dynes/cm^2. The height of this level varies with composition, and is also given in Table IV for each model atmosphere.

TABLE IV

Data on Lower Atmosphere and Mesopause for the different models

	Model	Lapse rate	Height of tropopause	Relaxation level	Height of mesopause	Density of mesopause	Temperature at mesopause
(b)	50% A 50% CO_2 5 mb	5.30 °C/km	15.9 km	163 km	200 km	8.2×10^{12}	195 °K
(c)	60% CO_2 25% Ne 15% N_2	3.86 °C/km	21.5 km	172 km	280 km	12.0×10^{12}	200 °K
(d)	5 mb 100% CO_2 1 mb	4.32 °C/km	19.4 km	112 km	150 km	4.9×10^{12}	210 °K

Above this level, following CHAMBERLAIN (1962), the CO_2 emission in the vibration-rotation bands at 15 μ is assumed lost to space. The volumetric loss of energy L_v is given by BATES (1951):

$$L_v = \eta n(CO_2) N e^{-h\nu/kT} h_\nu, \tag{6}$$

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where η = deactivation coefficient for CO_2; $n(CO_2)$ = number density of CO_2; N = total particle number density; h_ν = quantum energy at 15 μ; T = temperature of CO_2.

With increasing altitude L_ν decreases as the product of the number densities, n and N. At the level of the mesopause the total amount of energy radiated from above this height equals the solar ultraviolet energy which is conducted downward from the thermosphere. The temperature gradient at the mesopause is therefore zero. In order to compute the height and number density of individual constituents of the mesopause we need to know the density distribution below. This problem is complicated by the fact that CO_2 is partly photo-dissociated within this region, and a layer of O_2 may form which in turn may screen CO_2 from the solar ultraviolet radiation. MARMO and WARNECK (1961) have studied a similar problem for Mars. Following their procedure we first obtain the photochemical equilibrium distribution with height for CO_2, O_2, O and CO in this region. Using these numbers we then derive the height, density and temperature at the mesopause for each model. The results are shown in Table IV.

6.3. THERMOSPHERE AND EXOSPHERE

The temperature distribution above the mesopause is determined by the heat conduction equation,

$$\frac{\mathrm{d}}{\mathrm{d}z}\left(\kappa \frac{\mathrm{d}T}{\mathrm{d}z}\right) = Q - L, \tag{7}$$

where Q is the amount of solar ultraviolet and X-ray energy absorbed per unit volume, L is the energy radiated per unit volume, and κ is the thermal conductivity.

The input of the solar energy in the thermosphere in each model atmosphere is mainly due to absorption of ultraviolet radiation by CO, O and by Ne or A. This is because CO_2 is mostly dissociated at lower levels, near the mesopause. It is also assumed that above the mesopause diffusive separation occurs and the density distribution of each constituent is determined by its own scale height. Because of this circumstance, the solar energy deposited in the thermosphere is due to the absorption of the ionizing radiation of wavelengths less than 900 Å. Not all of this solar energy goes into heating. WALKER (1964) has made a detailed study of the efficiency of heating at different altitudes in the earth's atmosphere. Using his results, we adopt a mean efficiency (ε) of 50 percent for the conversion of solar ultraviolet to heat. The other half is radiated away and/or carried to lower levels by downward diffusion. Therefore the expression for Q for a slowly rotating planet, is

$$Q = \tfrac{1}{2}\varepsilon \sum_{\mathrm{i}} n_{\mathrm{i}}(z) \int_{\lambda_1}^{\lambda_2} F_\infty(\lambda)\, \sigma_{\mathrm{i}}(\lambda)\, e^{-\tau(\lambda,z)}\, \mathrm{d}\lambda \tag{8}$$

where ε = efficiency of conversion of solar ultraviolet to heat; F_∞ = incident solar flux; τ = optical thickness; λ = wavelength; z = altitude; n_{i} = number density of the ionizing constituent; σ = ionization cross section.

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The radiative loss term, L, in Equation (7) represents the total radiation per unit volume emitted by the atmosphere that is lost to space. As such, this is valid for the atmosphere above some level where the optical thickness is small and the absorption of radiation by the atmosphere above is negligible. For the atmospheric models under consideration, the emission is mainly by CO_2 in the 15 μ vibration-rotation band, and CO in the 4.66 μ band. The emission by atomic oxygen at 63 μ is assumed negligible.

High up in the thermosphere, where the temperature reaches a maximum, the term L is dominated by emission of CO. The energy L_r radiated per unit volume by CO (4.66 μ band) is given by Bates (1951):

$$L_r = 0.26\,(kT)^2\,B^2\,P^2\,n(\mathrm{CO})\ \mathrm{eV/cm^3/sec}, \tag{9}$$

where B=rotation constant of CO; P=dipole moment of CO; T=temperature; n(CO)=number density of CO. From the above expression it is seen that emission due to CO increases as T^2 which results in the dominance of CO radiation at high altitudes.

The level at which CO emission becomes important depends on the mean optical thickness of the rotation lines. McElroy, L'Ecuyer and Chamberlain (1965) have suggested that the height at which the optical thickness of the lines equals unity is the level above which CO radiation is dominant. Below this level CO_2 is the principal radiator and L is given by L_v (Equation (6)). But for calculating the exospheric temperatures, L can be replaced by L_r in models (b), (c), and (d). As model (a) only consists of an argon atmosphere, $L=0$.

The exospheric temperature can now be determined by solving the integro-differential equation (7) by a convergent iterative process. The first order solution for the exospheric temperature, T_∞, is obtained from the following expression:

$$\frac{A}{p+1}\left[T_\infty^{p+1} - T_m^{p+1}\right] = \tfrac{1}{2}\varepsilon F \Delta\lambda H_i\left[0.5772 + \ln\tau_m + E_i(\tau_m) - \frac{H_{CO_2}}{H_i}\right] - C n_r(\mathrm{CO})\,T_r^2/H_r^2(\mathrm{CO})\left[1 + \frac{z_r - z_m}{H_r}\right] \tag{10}$$

In this equation, T_m=mesopause temperature; A=conductivity coefficient in $\kappa = AT^p$; p=exponent of the temperature dependence of κ; F=average incident solar flux over the spectral band $\Delta\lambda$; $\Delta\lambda$=900–44 Å (856 Å); τ_m=mean optical thickness of the atmosphere at the mesopause to ionizing radiation; $E_i(\tau_m)$=exponential integral; H_{CO_2}=scale height of CO_2 at the mesopause; H_i=scale height of the principal ionized constituent (O or Ne) in the vicinity of the level of maximum absorption; C=constant=0.26 $k^2B^2P^2 = 2.9\times10^{-23}$; z_m=mesopause altitude, and the subscript r denotes the level at which the optical thickness of the CO rotation lines equals unity.

If we assume that the level z_r is high enough to make $T_r \approx T_\infty$, then $H_r = kT_\infty/mg_r$, where g_r is the gravitational acceleration at $z=z_r$ and m now corresponds to the molecular weight of CO. Also, it is assumed that the level of maximum absorption of solar ionizing radiation occurs at an altitude high enough to make $H_i \approx kT_\infty/m_i g$, where m_i

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is the molecular weight of the principal ionizing particle. With these assumptions, we solve Equation (10) for the exospheric temperature T_∞ for each of the atmospheric models. The results are shown in Table V.

TABLE V
Exospheric temperature

Model	Composition	Exospheric Temperature
(a)	100% A	> 10000 °K
(b)	50% CO_2 50% A	1100 °K
(c)	60% CO_2 25% Ne 15% N_2	1050 °K
(d)	100% CO_2	1100 °K

As mentioned above, these temperature values have been calculated on the assumption that $H_i \approx H_r \approx H_\infty$. However, the uncertainty in the exact level at which cooling by CO becomes effective and the level at which the maximum absorption of energy due to ionization occurs is large, and results in a wide range of values for the exospheric temperature. Also, there is considerable uncertainty in the magnitude of the solar ultraviolet flux and in the exact value of the heating efficiency factor ε. With these limitations in mind, it is perhaps worthwhile to emphasize that, depending on the assumptions made, the exospheric temperature of Mercury for models (b), (c) and (d) can be as high as 1800 °K or as low as 800 °K.

6.4. Discussion

There are two main conclusions that can be drawn from the results of Section 6.3:

(1) A pure argon atmosphere cannot be stable against gravitational escape. At $T_c > 10000$ °K, argon will escape in a relatively short time.

(2) If the observations of Moroz are correct and CO_2 is present in the atmosphere of Mercury, then enough CO is produced in the upper atmosphere by photodissociation so that the exospheric temperature in each of the three models (b), (c) and (d) will be in a range that will still permit an atmosphere that is stable against thermal escape. The large margin of error, 800 °K to 1800 °K, is mainly because of the uncertainty in the value of the solar ultraviolet flux, the efficiency factor, and the scale heights at which CO emission becomes dominant and the absorption of ionization energy is maximum.

It is interesting to note that despite the large differences in the molecular weight and thermal conductivity of argon and neon, models (b) and (c) give approximately the same temperature at the top of the atmosphere. This result is brought about because of the basic assumption that above the mesopause the atmosphere is in diffusive equilibrium and therefore, in all three models (b), (c) and (d), at the top

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of the atmosphere, atomic oxygen becomes the dominant constituent and the principal absorber of the ionizing radiation. Also, at high temperatures, the emission by CO is so intense that very little energy is available to be conducted down, and therefore the thermal conductivity of gases does not play an important role in determining the exospheric temperature. Only the temperature gradient in the thermosphere is affected, which in turn modifies the height of the exosphere. However, the changes are small and will not considerably alter our calculations of the flux of escaping gases.

It may also be pointed out that these calculations of the exospheric temperature do not take into account any effect of the transport of heat by circulation in the upper atmosphere.

The implication of these results appears to be that if the actual temperature of the exosphere of Mercury is close to our calculated lower limit, i.e., ~ 800 °K, then models (b), (c) and (d) are stable against gravitational escape. In this case the observed CO_2 may be of primordial origin and the atmosphere may contain substantial proportions of neon. However, one cannot rule out the possibility of accumulating an atmosphere by outgassing also. If, on the other hand, the average exospheric temperature is higher than 800 °K but less than 1200 °K, then both O and Ne will escape and the atmosphere will be mainly CO and A. Molecules of CO may recombine to form CO_2, but the process is not presently understood. If the exospheric temperature were to be between 1300 and 1800 °K, then even CO would be lost, resulting in an atmosphere composed mainly of argon and of that amount of CO_2 which will be in equilibrium between outgassing from the interior and gravitational escape from the top of the atmosphere.

In summary, therefore, it is suggested that the presence of CO_2 in the atmosphere of Mercury is not only highly probable, but is an essential condition for any substantial amount of atmosphere to exist on the planet.

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References

ADAMS, W. S. and DUNHAM, T., Jr.: 1932, *Publ. Astron. Soc. Pacific* **44,** 243.
ANTONIADI, E. M.: 1934, *La Planète Mercure*, Paris, Gauthier Villars.
BATES, D. R.: 1951, *Proc. Phys. Soc. London* **B 64,** 805.
BAUM, R. M.: 1960, in *A Guide to the Planets* by P. MOORE, New York, Norton and Co., Inc., Plate II.
BAUM, R. M.: 1963, in *The Planet Mercury* by W. SANDNER, New York, MacMillan Co.
CAMERON, A. G. W.: 1963, Yale University Lecture Notes, 1962–63, compiled by W. D. ARNETT, C. J. HANSEN, and J. W. TRURAN.
CHAMBERLAIN, J. W.: 1962, *Astrophys. J.* **136,** 582.
CHAMBERLAIN, J. W. and HUNTEN, D.: 1965: *Rev. of Geophys.* **3,** 299.
COLOMBO, G.: 1965, *Nature* **208,** 575.
DOLLFUS, A.: 1953, *Bull. Soc. Astron. France* **67,** 61.

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1966SSRv....5..565R

DOLLFUS, A.: 1957, *Ann. Astroph. Suppl.* **4.**
DOLLFUS, A.: 1961, in *Planets and Satellites: Solar System* **III** (G. P. KUIPER and B. N. MIDDLEHURST, eds.), Chicago, University of Chicago Press.
DOLLFUS, A.: 1962, in *Handbuch der Physik* **54.**
FIELD, G.: 1964, in *The Origin and Evolution of the Atmospheres and Oceans* (P. J. BRANCAZIO and A. G. W. CAMERON, eds.), New York, J. Wiley and Sons, Inc.
GOODY, R. M.: 1964, *Atmospheric Radiation*, Oxford, Clarendon Press.
HARRIS, D. L.: 1961, in *Planets and Satellites: Solar System* **III** (G. P. KUIPER and B. N. MIDDLEHURST, eds.), Chicago, University of Chicago Press.
HINTEREGGER, H. E.: 1961, *J. Geophys. Res.* **66,** 2367.
HOWARD, W. E. III, BARRETT, A. H., and HADDOCK, F. T.: 1962, *Astrophys. J.* **136,** 995.
JEANS, J. H.: 1916, *Dynamical Theory of Gases*, New York, Dover Publication, 1954.
KELLERMAN, K. I.: 1965, *Nature* **205,** 1091.
KOZYREV, N. A.: 1964, *Sky and Telescope* **27,** 339.
KUIPER, G. P.: 1952, in *Atmospheres of the Earth and Planets* (G. P. KUIPER, ed.), Chicago, University of Chicago Press.
LOWELL, P.: 1902, *Mem. Amer. Acad. Arts and Sci.* **12,** 431.
LYOT, B.: 1929, *Ann. Obs. Paris, Section de Meudon* **8** (NASA translation TT F–187, July 1964).
LYOT, B.: 1943, *Astronomie* **67,** 3.
MARMO, F. F. and WARNECK, P.: 1961, Final Report, Contract Nas w-124, Geophysics Corporation of America, Bedford, Massachusetts.
MAYER, C. H.: 1961, in *Planets and Satellites: Solar System* **III** (G. P. KUIPER and B. N. MIDDLEHURST, eds.), Chicago, University of Chicago Press.
MCELROY, M. B., L'ECUYER, J. and CHAMBERLAIN, J. W.: 1965, *Astrophys. J.* **141,** 1523.
MCGOVERN, W. E., GROSS, S. H., and RASOOL, S. I.: 1965, *Nature* **208,** 375.
MINTZ, Y.: 1962, *Icarus* **1,** 172.
MOROZ, V. I.: 1963, *Astron. Tsirk (USSR)* **270.**
MOROZ, V. I.: 1965, *Soviet Astronomy – AJ* **8,** 882.
ÖPIK, E. J.: 1961, *J. Geophys. Res.* **66,** 2808.
PEALE, S. J. and GOLD, T.: 1965, *Nature* **206,** 1240.
PETTENGILL, G. H. and DYCE, R. B.: 1965, *Nature* **206,** 1240.
PETTIT, E.: 1961, in *Planets and Satellites: Solar System* **III** (G. P. KUIPER and B. N. MIDDLEHURST, eds.), Chicago, University of Chicago Press.
SCHIAPARELLI, G. V.: 1889, *Astron. Nachr.* **123,** 241.
SPINRAD, H., FIELD, G., and HODGE, P.: 1965, *Astrophys. J.* **141,** 1155.
SPINRAD, H. and HODGE, P.: 1965: *Icarus* **4,** 105.
SPITZER, L. Jr.: 1952, in *Atmospheres of the Earth and Planets* (G. P. KUIPER, ed.), Chicago, University of Chicago Press.
UREY, H. C.: 1959, in *Handbuch der Physik* **52.**
WALKER, J. C. G.: 1961, *Astrophys. J.* **133,** 274.
WALKER, J. C. G.: 1964, Dissertation, Columbia University.

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