Results and Interpretation of the S-Band Occultation Experiments on Mars and Venus

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ABSTRACT

The S-band occultation experiments have provided important observational data on the structure of the atmospheres of Mars and Venus. The results for Mars show a very cold region in the middle atmosphere where CO_2, the main constituent of the Martian atmosphere, will saturate and may condense. Also, at the surface boundary of Mars there appears to be a temperature discontinuity between the air and ground in the local afternoon and a temperature inversion in the atmosphere in the nightside of the planet. On Venus, the Mariner 5 results show that the diurnal variation of temperature is small not only in the lower atmosphere but also in the stratosphere, implying strong zonal mixing at those levels. Also, the amplitude data from the Mariner 5 experiment has provided evidence for the existence of cloud layer or layers in the lower atmosphere where the ambient temperatures range between 350 and 450K. Such high temperatures preclude water as the constituent of these clouds. As for the upper atmospheres of both Mars and Venus, the occultation experiments indicate that the ionospheres of these planets contain up to a factor of two *more* electrons than can be explained in terms of the presently accepted values of the EUV flux. At the same time the exospheric temperature of Mars appears to be as low as 350–450K, about 100K *lower* than the closest predicted value.

1. Introduction

The highly successful S-band occultation experiments carried out by the Mariner fly-by missions to Mars and Venus have provided our first information on the structures of both the ionospheres and lower atmospheres of these planets. The purpose of this paper is to summarize the results of these experiments, discuss their inherent uncertainties, and compare them with theoretical models of the Mars and Venus atmospheres.

The first radio occultation experiment was successfully performed by the Mariner 4 fly-by of Mars (Kliore *et al.*, 1965). The most significant result of this historic experiment was to provide the first observation of the ionosphere of another planet. Consequently, the subject of the aeronomy of Mars was instantly elevated to a status of scientific respectability. This experiment, along with simultaneous spectroscopic observations (Belton and Hunten, 1966), revealed that the atmosphere of Mars was composed primarily of CO_2 and that the surface pressure was in the range of 5–10 mb.

Since the Mariner 4 flight the atmosphere of Mars has been explored by the occultation technique at four more locations on the planet (Fig. 1) and the atmosphere of Venus over two points (Fig. 2). The principles of the experiment, the methods of data retrieval, and the derivation of atmospheric temperatures, pressures, and electron densities from these data have been discussed extensively (Fjeldbo and Eshleman, 1968; Fjeldbo *et al.*, 1971) and will not be considered here in any detail. However, a brief review of the steps in the data reduction and of the method of error analysis, followed by a discussion of the accuracy of derived physical characteristics of the atmosphere, is in order.

2. Data reduction and error analysis

The basic data of the occultation experiment from which the results discussed in this paper are derived are the Doppler residuals as a function of time just before entry into and just after exit from occultation of the spacecraft by the planet. The steps involved in obtaining the electron density and lower atmosphere temperature profile are as follows:

1) Integrating the Doppler residuals with respect to time to obtain the phase path change.
2) Computing the bending angle of the ray from the spacecraft to the receiving station in order that the straight line approximation to the phase path can be calculated.
3) Calculating the atmospheric refractivity N as a function of radial distance from the center of the planet.
4) Computing the electron densities n_e[cm^{-3}] from the negative refractivities in the upper atmosphere [$n_e = -(f^2/40.3\times 10^{12})N$, where f is the frequency of the S-band signal].
5) Assuming a composition for the atmosphere and converting the lower atmosphere refractivity profile into a density profile.

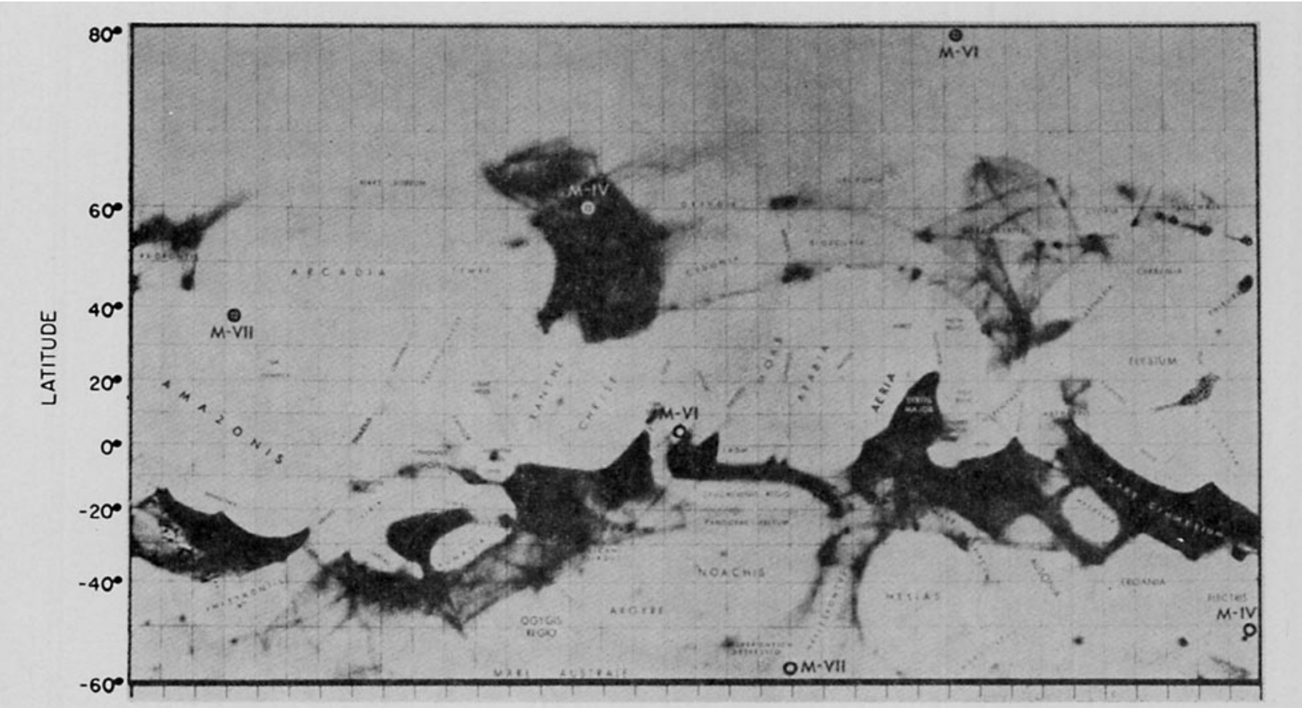

FIG. 1. Locations of the Mariner 4, 6, and 7 occultation points on the globe of Mars (after Kliore *et al.*, 1970).

6) Assuming the hydrostatic and perfect gas laws and integrating the density profile in the lower atmosphere to obtain the temperature and pressure distributions down to the surface of the planet.

In cases of relatively dense planetary atmospheres (Earth, Venus, Jupiter, etc.) the occultation experiment provides another important set of data which can be used to determine attenuation properties of the atmosphere at radio wavelengths. This information is obtained by monitoring the amplitude of the radio signal as it traverses the atmosphere. The decrease in the signal strength as a function of time can be translated into atmospheric attenuation (db km^{-1}) vs altitude (Kliore *et al.*, 1969). Such measurements from the S-band occultation experiment on Mariner 5 have

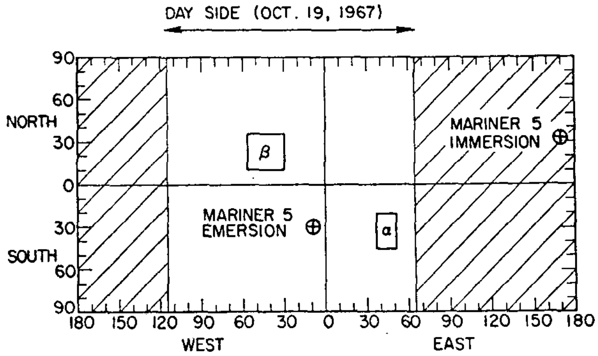

FIG. 2. Locations of the Mariner 5 occultation points on the globe of Venus (after Fjeldbo *et al.*, 1971).

provided important data on microwave attenuation in the lower atmosphere of Venus (Kliore *et al.* 1969; Rasool, 1970; Fjeldbo *et al.* 1971).

The derivation of physical properties of the Mars and Venus atmospheres from the Doppler residual data outlined above involves two types of errors: 1) systematic errors due to reasons such as oscillator drift and imperfect knowledge of the spacecraft trajectory, and 2) random errors due to the noise level of the data. In addition to these sources, errors in the interpretation of the data may result because of the assumption of global symmetry in the vertical distribution of both the neutral gases in the lower atmosphere and the electrons in the upper atmosphere. Also, the possible presence of electrons in the lower atmosphere may be yet another source of error in the derivation of neutral atmosphere densities from the refractivity data (Rasool *et al.*, 1970).

Systematic errors are manifested in non-zero values of the Doppler residuals corresponding to times when the spacecraft is far from the planet and atmospheric effects due to the planetary atmosphere cannot be present in the spacecraft telemetry signal. These errors are removed from the data by performing a polynomial regression analysis on the Doppler residuals over selected time periods in the pre-encounter (entry) and post-encounter (exit) periods. The resultant least-squares fitted bias curve is then extrapolated through the encounter period and subtracted from the complete set of Doppler residuals to obtain a set of corrected residuals (Fig. 3). Operations 1)–6) described earlier are then carried out on these corrected residuals to obtain physical properties of the planetary atmosphere.

As for the random error, the polynomial regression analysis yields the variance of the Doppler residual points to which the bias curve is fitted. This initial error will propagate through the calculations and will ultimately result in error limits on the derived atmospheric profiles. Errors in all quantities derived from the Doppler residuals will be related to the random error in the residuals themselves and must be evaluated by considering the series of mathematical operations performed on the residuals. It is assumed that the errors in the Doppler residual points are statistically independent and that uncertainties in other factors entering the analysis, such as wavelength of the telemetry signal and spacecraft position, are either negligible compared to errors in the residuals or are systematic and hence removed in the data reduction.

According to the analysis of Stewart and Hogan (1972), both the systematic and the random errors in the data result in relatively small uncertainties in derived physical properties of the lower atmosphere near the ground. The random errors become large, however, in the middle atmosphere where the refractivities are small.

3. Results

In this section we will summarize the values of the atmospheric parameters and their attendant uncertainties that have been derived for Mars from the Mariner 4, 6 and 7 data and for Venus from the Mariner 5 data. The discussion of these results in conjunction with theoretical models of the atmospheres of Mars and Venus will be presented in the next section.

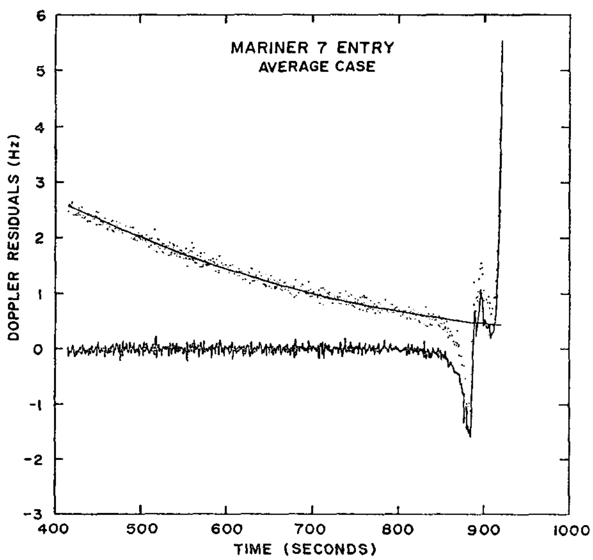

FIG. 3. Raw residuals (dots), third-order polynomial fit, and corrected residuals for the Mariner 7 entry data averaged from three receiving stations.

TABLE 1. Surface pressures and temperatures on Mars from Mariner occultation experiments.

	Latitude	Longitude	Local time	χ (deg)	Surface pressure (mb)	Surface temperature (°K)
Mariner 4 entry	50.5S	177.0E	1300	67	4.9	160
Mariner 6 entry	3.7N	4.3W	1545	57	5.4	254
Mariner 7 entry	58.2S	30.3E	1430	56	5.0	221
Mariner 4 exit	60.0N	34.0W	0030	104	7.6	(240)
Mariner 6 exit	79.3N	87.1E	2210	107	8.5	152
Mariner 7 exit	38.1N	148.3W	0310	130	8.0	209

a. *Mars: Surface*

Table 1 summarizes the values obtained by Mariners 4, 6 and 7 for the atmospheric temperature and pressure near the surface at six different locations on Mars shown in Fig. 1. Mariner 4 flew by Mars on 15 July 1965, when the solar declination at Mars was $+11°$. Entry into occultation occurred near 50S latitude corresponding to Southern Hemisphere winter, while exit occurred near 60N latitude corresponding to Northern Hemisphere summer. Mariner 6 flew by Mars on 31 July 1969, when the solar declination at Mars was $-8°$. Entry occurred near 4N (early fall) and exit near 79N. Mariner 7 flew by five days later on 5 August 1969. Entry occurred near 58S (early spring) and exit near 38N.

The values presented in Table 1 have been calculated assuming the composition of the Martian atmosphere to be 100% CO_2. Since the observed refractivity scale height essentially fixes the ratio of temperature to mean molecular mass, the effect of the presence of lighter gases in the atmosphere, such as N_2 or Ne, would be to lower the temperature and increase the pressure values shown in the table. However, for a given composition a comprehensive analysis of both systematic and random errors, as discussed in the previous section, suggests an uncertainty of about $\pm$6K in temperature and $\pm$0.5 mb in pressure.

b. *Mars: Lower atmosphere*

The atmospheric temperature profiles above these occultation points are shown in Figs. 4 and 5. Fig. 4 shows the three dayside profiles and Fig. 5 the three nightside. The Mariner 4 profiles are those published by Fjeldbo and Eshleman (1968) while the Mariner 6 and 7 profiles were obtained by Hogan *et al.* (1971). The solid line representing the Mariner 6 and 7 temperatures is an average obtained from 300 reductions of the Doppler residual data using third-order polynomial fits to different sets of pre- or post-encounter points to remove the systematic error in each case. The shaded region about the Mariner 7 profiles represents the standard deviation in the average temperature values. The corresponding standard deviations in the Mariner 6 profiles are too small to be shown as shaded regions over most of the altitude range for which temperatures were derived. The motivation for analyz-

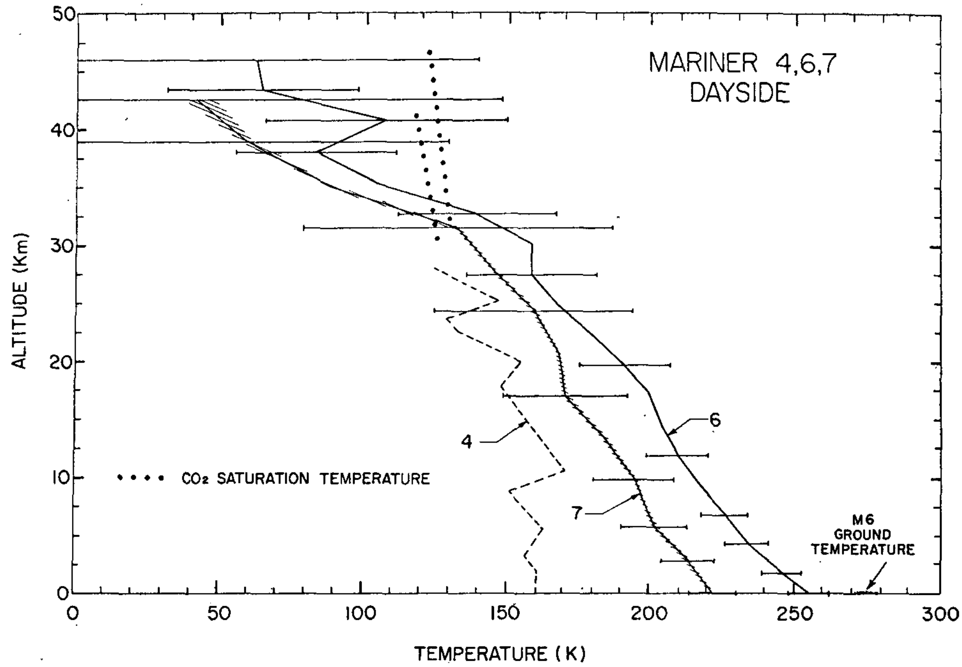

FIG. 4. Lower atmosphere dayside temperature profiles for Mars obtained from the Mariner 4, 6 and 7 occultation data. See text for a discussion of the error bars.

ing a large number of different cases is to obtain an estimate of the error involved in extrapolating the bias curve through the residual data pertaining to the atmosphere. Clearly, the scatter in the temperatures obtained in different data reductions is small and the extrapolation of the fitted polynomial through the encounter period does not appear to be a serious source of error.

The error bars shown on the Mariner 6 and 7 profiles are the standard deviations in temperature resulting

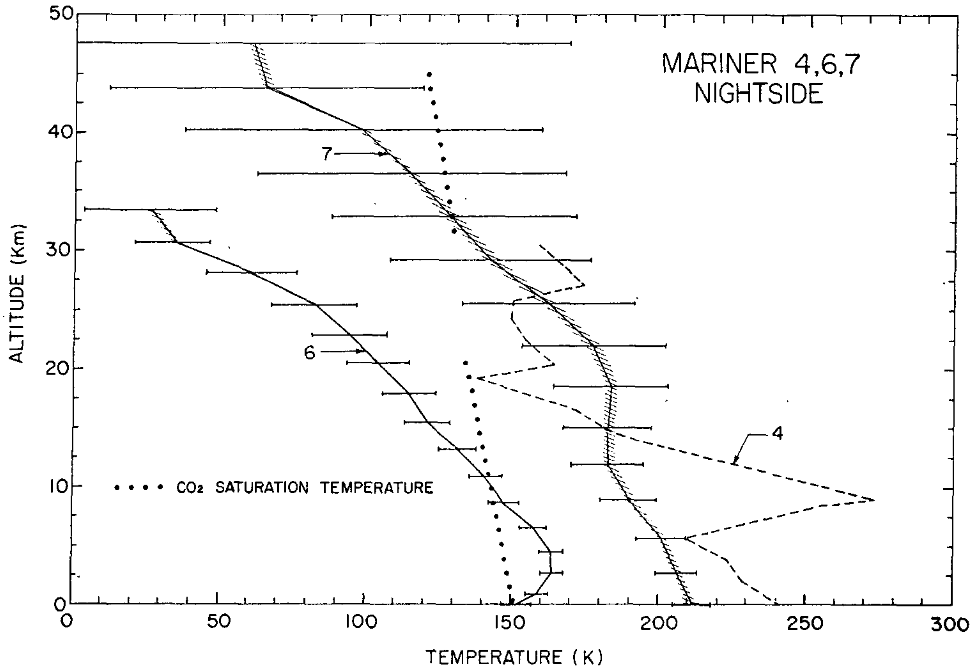

FIG. 5. Lower atmosphere nightside profiles for Mars obtained from the Mariner 4, 6, and 7 occultation data. See text for a discussion of the error bars.

from the noise level of the residual data and are typical of the random errors which accumulate in any single reduction of the data (Stewart and Hogan, 1972). If we assume that the procedure described above gives a realistic estimate of the systematic error involved in the data reduction, then the random error in temperature exceeds the systematic error at all altitudes. Near the ground these errors are about ± 5 to ± 7K, being somewhat larger in the Mariner 7 cases. The magnitude of the temperature errors generally increases with altitude.

The Mariner 6 and 7 profiles shown in Figs. 4 and 5 are similar to those published earlier by Rasool *et al.* (1970). The principal difference is in the Mariner 6 exit profile which is now about 10K colder at all altitudes than that obtained in the earlier analysis. The results of Rasool *et al.* were based on a single reduction of the residual data and employed a different bias removal technique. Since the present results are based on more extensive analyses of the data they should be regarded as more reliable.

c. Venus: Lower atmosphere

Figs. 6 and 7 show the night- and dayside temperature profiles obtained for the Venus atmosphere by Mariner 5 (Fjeldbo *et al.*, 1971), assuming a composition of 95% CO_2, 5% N_2. Fjeldbo *et al.* estimate that the Mariner temperature profiles are accurate to ± 1K below 60 km altitude. In Fig. 6 we have plotted the *in situ* temperature measurements made by Veneras 4, 5 and 6 (Avduevsky *et al.*, 1970). The Mariner profile has been extrapolated adiabatically to the ground and the value agrees with the recently measured temperature of 747 ± 20K at the surface of Venus by Venera 7 (Avduevsky *et al.*, 1971). The pressure at the surface as derived from the Soviet experiment is 90 ± 15 atm. Using this value and assuming an adiabatic atmosphere, we have also plotted in Figs. 6 and 7 the approximate pressure levels which correspond to various altitudes in the nightside atmosphere of Venus. As all Soviet measurements were made on the nightside, the Mariner profile is the only datum we possess for the temperature structure on the dayside.

Another important datum on the lower atmosphere of Venus obtained by the occultation experiment is the evidence of strong attenuation of S-band signal at all altitudes below 52 km, both on the nightside and dayside of the planet. This feature was first described by Kliore *et al.* (1969), elaborated on by Rasool (1970), and recently confirmed by Fjeldbo *et al.* (1971) who made a more comprehensive analysis of occultation data at three different frequencies.

The atmospheric loss coefficient at S band, as derived by Fjeldbo *et al.* for altitudes in the Venus atmosphere between 30 and 60 km, is plotted as an insert in Figs. 6 and 7. This loss coefficient has been derived after accounting for the atmospheric "defocusing" of the signal caused by the vertical density gradient in the atmosphere. It therefore corresponds to actual attenuation of the microwave radiation by the atmosphere and consequently indicates the presence of some absorbing or scattering material at these levels in the Venus atmosphere. As far as it can be evaluated from the data published by Fjeldbo *et al.*, the uncertainty in these numbers for the loss coefficient is $\pm 0.1\times 10^{-2}$ db km^{-1}.

d. Mars and Venus: Upper atmosphere

The basic result of the occultation experiment for the upper atmospheres of Mars and Venus is the distribution of electron density with altitude. The ionospheric profiles for Mars obtained by Mariners 4, 6 and 7 are shown in Fig. 8. All three correspond to the dayside of the planet. The Mariner 4 result is that of Fjeldbo and Eshleman (1968), while the Mariner 6 and 7 profiles are from Hogan *et al.* (1971). The maximum electron density at the time of the Mariner 4 flight (near solar minimum) was $9\pm 1\times 10^4$ cm^{-3} and occurred near 120 km. Four years later, during a period of intermediate solar activity, Mariners 6 and 7 observed increased electron densities, between 1.6 ± 0.1 and $1.7 \pm 0.1\times 10^5$ cm^{-3} near 135 km. The Mariner 6 and 7 electron densities shown in Fig. 8 are average values obtained from 300 reductions of the residual data as explained earlier in connection with the lower atmosphere temperature profiles. The standard deviations in these average values are much smaller than the random errors noted above.

The Mariner 5 flight past Venus on 19 October 1967, at a time of intermediate solar activity, observed a maximum electron density of $5.5\pm 0.5\times 10^5$ cm^{-3} near 142 km altitude on the dayside of the planet and a maximum of about 2×10^4 cm^{-3} at the same altitude on the nightside [Fjeldbo and Eshleman, 1968; the summary shown in Fig. 9 has been adopted from Eshleman (1970)]. The dayside ionization profile exhibits a plasmapause, or sudden drop in electron density, near 500 km. The error estimate on the dayside maximum electron density is that of Kliore *et al.* (1969).

In addition to the major ionization peak all the Mariner flights have observed a secondary electron density ledge at lower altitudes as evidenced in Figs. 8 and 9. In the Mars ionosphere this ledge was observed 25 km below the primary maximum by Mariners 4, 6 and 7. The Mariner 5 observed the secondary ledge 15 km below the primary maximum in the dayside Venus ionosphere.

If the major electron density maxima in the Mars and Venus ionospheres are interpreted as F1 layers in photochemical equilibrium, then the plasma scale heights above the peak can be used to infer an exospheric temperature. At the time of the Mariner 4 flight the exospheric temperature obtained in this way was $\sim$275K. The Mariner 6 and 7 plasma scale heights yield

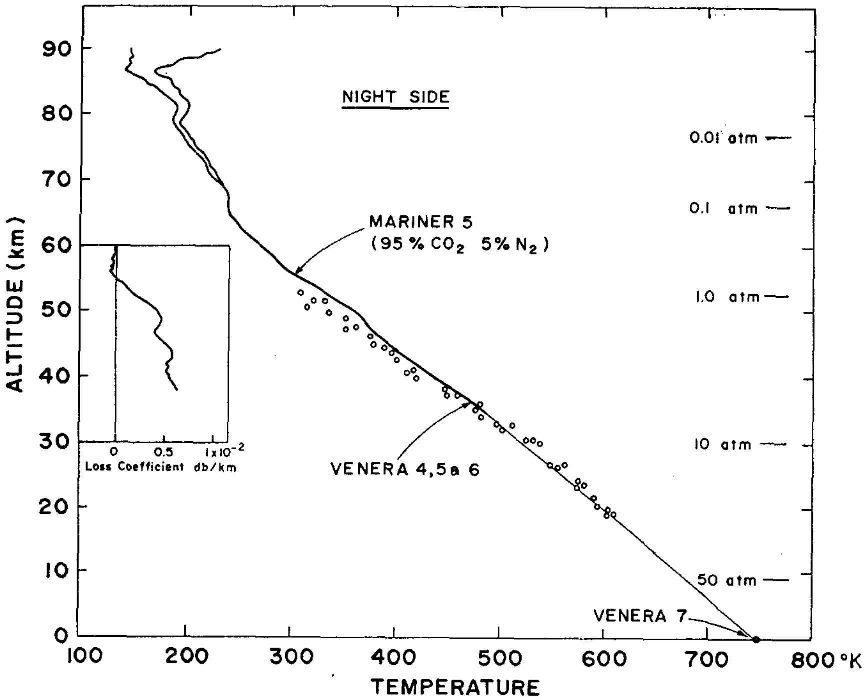

FIG. 6. Nightside temperature profile in the lower atmosphere of Venus. The insert shows the S-band loss coefficient between 30 and 60 km. The circles are the Venera 4, 5 and 6 data points.

exospheric temperatures of 388±54K and 425±35K, respectively (Hogan *et al.*, 1971). The plasma scale height above the electron density maximum in the Venus ionosphere shows a greater increase with altitude than does that in the Martian ionosphere, but it is matched quite well by thermal structure models having an exospheric temperature near 650K (McElroy, 1969; Hogan and Stewart, 1969).

4. Discussion

a. Mars: Surface pressure

The values for the atmospheric pressure on the surface of Mars at six locations on the planet (Fig. 1) range between 4.9 and 8.5 mb (Table 1). The observed differences in total pressure at various points are real and almost certainly due to the topography of the surface of Mars; the lower pressure values correspond to elevated regions while the higher pressures indicate depressed areas on the surface. Because the pressure scale height in the lower atmosphere of Mars is ~10 km, the observed pressure extremes would correspond to altitude variations on the surface of Mars of as much as 6 km. These results are consistent with those recently obtained by radar (Pettengill *et al.*, 1969), and with the observation of the differences in abundances of CO_2 over the disc of Mars both from ground-based measurements (Belton and Hunten, 1969) and from Mariners 6 and 7 (Herr *et al.*, 1970). A comprehensive discussion of the up-to-date results on the topography of Mars obtained by all the relevant techniques has recently been published by Belton and Hunten (1971).

b. Mars: Lower atmosphere

The Mars temperature profiles shown in Figs. 4 and 5 contain several features of interest.

In all the thermal profiles derived from the Mariner 6 and 7 occultation experiments, temperatures in the middle atmosphere fall below the value at which CO_2 would saturate. The possible condensation of CO_2 in the middle atmosphere of Mars is the most pervasive single feature of the temperature profiles deduced from the occultation data and occurs regardless of the details of the raw data adjustment. If the occultation data alone suggested CO_2 saturation in the middle Martian atmosphere, then this result would have to be viewed with some skepticism since the relative error in refractivity values are largest in this region where the refractivities are small and the corresponding errors in temperature are therefore large (Figs. 4 and 5). The fact that every temperature profile shows such an effect would imply that it might be real, although an alternative explanation, in terms of the possible existence of a Martian D region ionosphere, was noted by Rasool *et al.* (1970). There are, however, indications other than the occultation data which suggest that CO_2 condensation does occur in the middle atmosphere of Mars.

The infrared spectrometer aboard both Mariners 6 and 7 recorded a reflection at 4.3 μ at altitudes well

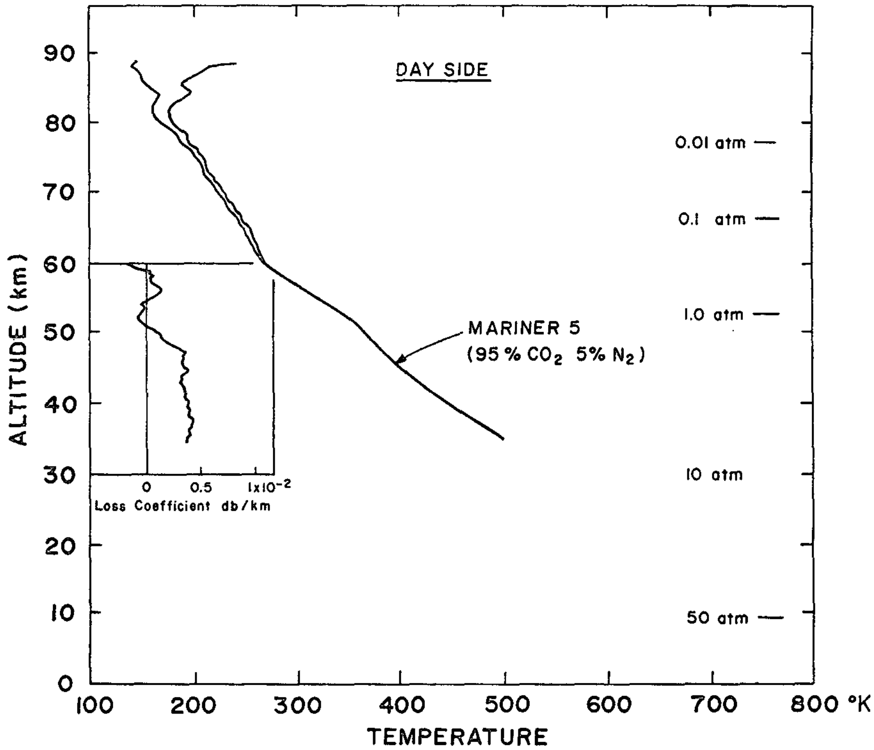

FIG. 7. Dayside temperature profile in the middle atmosphere of Venus. The insert shows the S-band loss coefficient between 30 and 60 km.

above the planetary surface (Herr and Pimentel, 1970). For the second Mariner 7 bright-limb crossing the experimenters estimate the altitude of the reflection to be 25±7 km. This reflection spike is due to the ν_3 fundamental of CO_2 and indicates the presence of condensed CO_2 in the Mars middle atmosphere.

A third line of evidence bearing on the possibility of CO_2 condensation comes from the Mariner 6 and 7 television pictures (Leovy *et al.*, 1971). Thin haze layers ranging from 5–50 km are observed above the limb of Mars in several pictures. Some of these layers are high enough in the atmosphere so that the scattering could be due to solid CO_2 (although as noted by Leovy *et al.* the lower altitude layers seem to require another mechanism).

Prior to the Mariner 6 and 7 experiments no theoretical models of the Martian lower atmosphere predicted condensation of CO_2. Recently, Gierasch (1971) has shown that if low dissipation is assumed in the Martian atmosphere the kinetic energy generated by

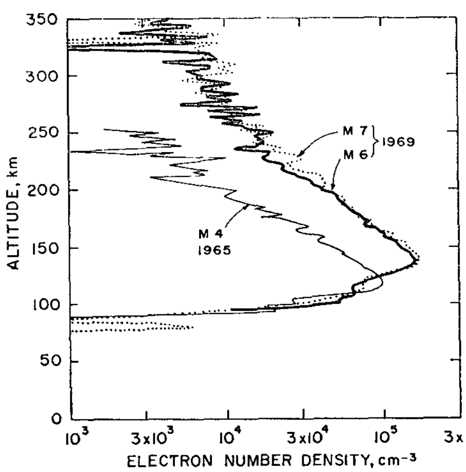

FIG. 8. The Martian ionosphere as observed during the Mariner 4, 6 and 7 occultation experiments.

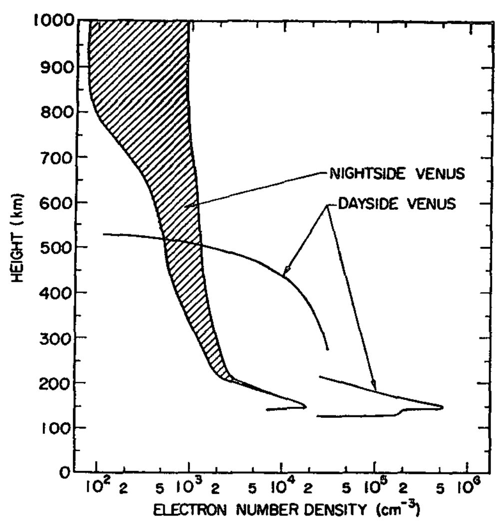

FIG. 9. The Venus ionosphere as observed during the Mariner 5 occultation experiment (adapted from Eshleman, 1970).

buoyancy would not be dissipated locally, but by an overshooting of turbulent motions into what would otherwise be a stable radiative regime. This would raise the tropopause height and result in substantially lower temperatures in the middle atmosphere. At low latitudes the temperature would be driven low enough for CO_2 to condense. Observations (Figs. 4 and 5) indicate, however, that condensation occurs at high as well as low latitudes and that the temperature gradient in about the first 20 km of the atmosphere is sub-adiabatic. Neither of these features is predicted by the theory of Gierasch and we must conclude at present that there is no satisfactory explanation for the observed thermal structure of the lower atmosphere of Mars.

The temperature of the atmosphere near the surface at the Mariner 6 entry occultation point is 254±7K while the *ground* temperature obtained from the infrared radiometer for the same region is about 275±5K (Neugebauer *et al.*, 1969, 1971). These results thus indicate an air-ground temperature discontinuity of ~20K in the late Martian afternoon (Fig. 4). The existence of a temperature discontinuity between the atmosphere and the surface of Mars was predicted by Gierasch and Goody (1968) on the basis that the Martian atmosphere is optically thin and is radiatively decoupled from the ground. According to these authors, turbulent transfer is the dominant process for heating the lower atmosphere and for the estimates of this parameter used in their calculations Gierasch and Goody obtain a maximum temperature discontinuity of +65K for the early afternoon and −40K for the early morning hours. The present results confirm the predictions of Gierasch and Goody that a temperature discontinuity exists at the atmosphere-surface boundary of Mars. However, the observed value of +20K, in contrast with the predicted value of +40K for 1600 hours Martian local time (MLT) may be taken as an indication that the magnitude of turbulent transfer of heat at the Martian boundary layer is higher than assumed by these authors.

In the same context, the Mariner 6 exit profile, corresponding to 2200 MLT, exhibits a strong temperature inversion near the ground. The maximum temperature occurs near 4 km and is about 15K higher than the temperature near the surface. However, neither the Mariner 4 nor the Mariner 7 nightside profiles exhibit such an inversion. According to the time-dependent calculations of Gierasch and Goody (1968), all the nightside profiles of Fig. 4 should exhibit an inversion. The Mariner 6 exit profile thus agrees with these predictions, but the lack of an inversion in the Mariner 4 and Mariner 7 exit profiles cannot be interpreted as a disagreement with theory. The error bars derived for the Mariner 7 exit profile would permit a mild inversion in this case and although no complete error analysis exists for the Mariner 4 profile it is likely that an inversion could be present.

c. Venus: Lower atmosphere

The most interesting features of the Venus temperature structure shown in Figs. 6 and 7 are the overall similarity of the dayside and nightside profiles, and the distinct changes in the temperature gradients which occur at about the same altitudes on both sides of the planet.

Because the Venus atmosphere is so massive ($p_s \approx 90$ atm), has a high heat capacity, and is optically thick both in the infrared and in the visible, most of the theoretical investigations have suggested that the diurnal variation of temperature in the lower atmosphere should be small, of the order of 5K or less [see, for example, the review by Goody (1969)]. The lower parts of the temperature profiles shown in Figs. 6 and 7 are the first observational evidence corroborating such calculations.

The more surprising aspect of these thermal profiles is that the temperatures in the stratosphere of Venus, at altitudes as high as 70 km and pressure levels as low 10–100 mb, do not show any significant diurnal variation. If, as is generally believed, the atmosphere at this level were in radiative equilibrium, one would expect day-to-night differences of the order of 70K at the 40-mb level (Bartko and Hanel, 1968). However, the observed amplitude of the diurnal oscillation of the temperature is probably less than 5K. This is an extremely significant result and has direct bearing on the possible existence of a strong zonal circulation in the Venus stratosphere. In fact, Gierasch (1970) has shown that a strong zonal flow of as much as 100 m sec^{-1} at ~40 mb would reduce the amplitude of the diurnal change in temperature to <5K. The result is independent of the details of his theoretical model for the source of the wind, and the small observed change therefore appears to be evidence that this level in the atmosphere corresponds to the now much discussed altitude of the ultraviolet clouds which seem to rotate with a period of 4 days (Boyer and Carmichel, 1965; Smith, 1967) corresponding to a velocity of ~100 m sec^{-1}.

The second important feature of these profiles is the distinct change in the temperature gradient at two levels in the atmosphere, 47 and 60 km, both on the dayside and nightside of the planet. The change in gradient from adiabatic to sub-adiabatic at the 60-km level probably defines the top of the convective region on Venus. However, at 47 km the fluctuations in the gradient are almost certainly related to the presence of a cloud layer at this altitude. The relevant observation in this regard is the measurement of atmospheric loss coefficient plotted as an insert in Figs. 6 and 7. In the atmosphere *above* 55 km the tangential attenuation of the S-band signal is less than 10^{-3} db km^{-1}. However, *below* this altitude substantial attenuation is observed with distinct peaks at 47 and 40 km on the nightside and a more uniform structure on the dayside. The temperature at the 47-km level is ~370K and, as pointed out by Rasool (1970), if these peaks are due to clouds they cannot be composed of water. An alternative suggestion of the possible condensation of mercury compounds at these temperatures has been made by Lewis (1969) and shown to match the observations by Rasool (1970).

The new results of the occultation experiment on the existence of fluctuations in the temperature gradient at the same level, where S-band attenuation goes through a peak value, lend weight to the possibility that a dense cloud layer exists at these low altitudes. However, because of the high ambient temperatures at this level and the extremely low abundance of water vapor in the upper atmosphere of Venus ($\sim$0.4% at the 300K level), it appears highly unlikely that these clouds could be composed of liquid water. Some other condensable vapor has to be invoked and if it is a mercury hallide or sulphide, the observed "loss coefficient" can be matched by a cloud layer containing 200 particles cm^{-3} of diameter 20 μ provided their dielectric loss tangent is close to 0.001, as appears to be the case (Westphal, W. B., private communication).

d. Venus and Mars: Upper atmosphere

The primary electron density maxima observed in the Mars upper atmosphere are now generally accepted to be F1 layers, though there is still some controversy as to whether they are in strict photochemical equilibrium (R. W. Stewart, 1971) or are modified via interaction with the solar wind (Cloutier *et al.*, 1969). If the UV fluxes reported by Hall and Hinteregger (1970) are assumed to be the only source of ionization, then the observed electron density profiles exhibit excess ionization of up to a factor of 2 over theoretical models (R. W. Stewart, 1971; Donahue, 1971).

The electron density profile in the dayside upper atmosphere of Venus seems to be similar in most respects to those observed on Mars. It appears to be an F1 layer and is in strict photochemical equilibrium (i.e., not modified by solar wind interaction). There is also about a factor of 2 more ionization in the Venus ionosphere than can be accounted for in terms of the EUV fluxes of Hall and Hinteregger (1970). Venus, unlike Mars, exhibits an ionosphere maximum on the nightside of the planet. McElroy and Strobel (1969) have suggested that the nightside ionosphere can be accounted for by transport of light ions such as He^+ and H_2^+ from the dayside and subsequent chemical reactions with CO_2 (and possibly N_2).

The secondary electron density maxima observed 25 km below the primary peak on Mars and 15 km below the major peak on Venus are analogous to the Earth's E region. It appears naturally in theoretical models of the Mars and Venus ionospheres (R. W. Stewart, 1971) and is a consequence of ionization caused by soft x rays. The existence (though not magnitude) of this secondary ledge in theoretical models does not depend on special assumptions regarding the composition of the neutral atmosphere. The smaller separation between primary and secondary electron density maxima in the Venus ionosphere is due to the larger value of g for that planet.

Presently, there does not appear to be any adequate theoretical framework within which the set of observations we now have for the Mars and Venus upper atmospheres can be understood. Both the plasma scale heights deduced from the occultation data (Hogan *et al.*, 1971) and the neutral scale heights deduced from UV spectrometer data (A. I. Stewart, 1971) indicate substantially lower exospheric temperatures on Mars than are obtained from the thermal structure model of McElroy (1969). The fact that the plasma scale height is about twice the neutral scale height also implies the absence of any solar wind alteration of the plasma distribution. The pre-Mariner 6 and 7 models of Stewart and Hogan (1969), based on EUV fluxes from Hinteregger *et al.* (1965), also predicted exospheric temperatures which appear to have been about 100K too high. Use of the solar EUV fluxes of Hall and Hinteregger (1970) would result in exospheric temperatures for Mariner 6 and 7 conditions in better agreement with observations, but only at the expense of the good agreement previously obtained for Mariner 4 and 5 conditions on Mars and Venus (Donahue, 1971).

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