COMPARATIVE CHARACTERISTICS OF THE IONOSPHERES OF THE PLANETS OF THE TERRESTRIAL GROUP: MARS, VENUS, AND THE EARTH

K. I. GRINGAUZ and T. K. BREUS
Radiotechnical Institute of the Academy of Sciences of the U.S.S.R., Moscow, U.S.S.R.

(Received 16 September, 1969)

Contents:

The first attempts to explore the vicinity of the planets of the terrestrial group (Mars and Venus) by space probes were made rather long ago (Venera 1, 1961; Mars 1, 1962, and Mariner 2, 1963).

From the very beginning the ionospheric parameters were thought to be interesting planetary characteristics (Gringauz, 1961). Intuitively it could be assumed that Venus, which has a dense atmosphere and is closer to the sun than the earth, has a denser ionosphere than the earth, while Mars, with its tenuous atmosphere, and which is farther from the sun could correspondingly have a less intense ionosphere.

Radioastronomic and radar investigations of Venus and determinations of the brightness temperatures of the Venusian radio emission at 0.4–40 cm led to the hypothesis that the electron density in the ionosphere of Venus could exceed the terrestrial ionospheric electron density by a factor of $\sim 10^3$ (Jones, 1961; Danilov and Jatcenko, 1964, 1965).

Martian ionospheric models were extremely controversial.

The lack of data about planetary atmospheres and magnetic fields was, naturally, the main reason for the difficulties in constructing models of ionospheres of planets. Recent experiments made it possible to measure the characteristics of the Martian and Venusian ionospheres for the first time.

The present paper is a brief review of the state of investigations of the ionospheres of planets of the terrestrial group. The following items are considered – the techniques of investigating planetary ionospheres, experimental results available and their interpretation, and comparison of these results.*

1. Methods of Investigating Planetary Ionospheres by Means of Space Vehicles

The characteristics of planetary ionospheres can be measured by plasma probes on

* In preparing this review, data available before July 1969 were used.

Space Science Reviews **10** (1970) 743–769.

vehicles approaching the surface of planets or on artificial satellites of planets, and by means of radio sounding of the planetary ionospheres from the planet's satellites – orbital ionospheric stations. They can be determined also by studying radio wave propagation between the earth and a spacecraft when the radio waves pass through the planetary atmosphere and ionosphere, in particular, during the radio-occultation of the spacecraft by a planetary atmosphere.

The use of the method of pulse sounding with the aid of an artificial orbital ionospheric station at a sufficiently close distance (a method now widely used for exploring the terrestrial ionosphere) permits systematic observations of the space- and time-dependent variations of the electron density in the planetary ionosphere, but involves great difficulties: the stations should operate on a wide wavelength range (since the critical frequencies of the relevant ionospheres are initially unknown) which means a still greater weight of the receiving-transmitting equipment than for earth-orbiting ionospheric stations and hence limits the possibility of simultaneous other scientific measurements on the same vehicle.

While using probe methods one may limit oneself to a fairly light instrumentation; probes may be combined easier with the whole complex of other physical studies. Starting with Sputnik III, probes for terrestrial ionospheric studies, in particular charged particles traps, have been widely used. At present there are hundreds of such instruments, installed on space vehicles of various countries, for conducting experiments aimed at studying charged particles in the earth's ionosphere. The designs of traps for exploring ionospheres of other planets could, e.g., be similar to those used for the earth's ionosphere-exploring satellites of the Electron type (Bezrukikh and Gringauz, 1965). The sensitivity range of such a device can be chosen in accordance with the supposed values of the charged particle densities in the ionosphere under investigation.

This method requires also the installation of special instrumentation in a space vehicle launched to a planet.

The method based on studies of radio waves, propagating between a space vehicle and the earth passing through the atmosphere of a planet, is the cheapest and easiest to realise because in its simplest variety it does not require special instrumentation, and one can use radio waves emitted from the spacecraft for telemetry communication or for the determination of the trajectory.

Figure 1 represents a schematic picture of the radio occultation of a spacecraft by a planet, which permits one to easily understand the main principles of this method.

When radio waves reach the atmosphere or ionosphere of the planet, their velocity of propagation varies as compared with the speed in the interplanetary medium (which is close to the velocity in vacuum), since the refraction indexes of neutral and ionized media differ from unity. The altitude gradient of the refraction index in the atmosphere and ionosphere produces refraction. At appropriate wavelengths these two phenomena may cause a significant change of the phase path. Time-dependent variations of the refraction indexes in the interplanetary medium, the earth's troposphere and ionosphere contribute further to an increase of the phase path. Thus the increase of the phase path measured during the occultation of a space vehicle by a planetary atmosphere is due to the motion of the spacecraft and to changes in the refraction index on the path of the radio waves. If the contribution due to the increase in the distance between the space vehicle and the ground-based receiving point (which would take place if the refraction index along the entire path were unity) and the contribution due to changes in the refraction index in the earth's ionosphere and interplanetary

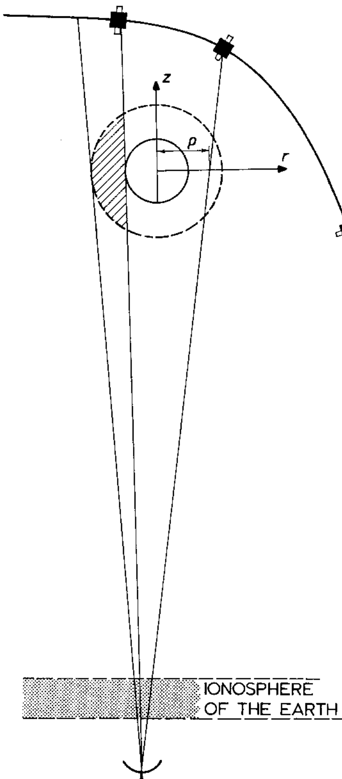

Fig. 1. Schematic representation of the spacecraft's radio-occultation by the planet. The z-axis passes through the centre of the earth and the planet. ϱ is the distance from the z-axis to the radio-beam.

space, are subtracted from the total increase of the phase of the received radio waves measured on the earth, then the experiment may under certain assumptions yield the height profile of the refraction index in the planetary atmosphere. Subsequently, in that altitude range where the refraction index is less than unity, one can determine the altitude dependence of the electron density. In the lower part of the atmosphere of a planet the refraction index exceeds unity. This section of the refraction index profile is used for determining the characteristics of the planetary neutral atmosphere. For instance, if the composition of the neutral atmosphere is known, it is possible to calculate, from this profile the neutral particle density profile in the lower atmosphere (since the refraction index for a mixture of gases is equal to the sum of the refraction indexes of all components and is for each component proportional to its density). This enables one to obtain as well pressure and temperature profiles.

The experiment with the radio occultation of a spacecraft by a planet can be realized in various modifications – it is feasible to use radio waves at one frequency or at two or more frequencies.

If one frequency is used one cannot unambiguously separate the contributions of the neutral and charged particles (mutually compensating for each other) to the relevant phase path variation, and it is necessary to assume that a lower planetary atmosphere does not contain a sufficient amount of charged particle densities, to essentially change the value of the refraction index under determination. In fact, the overall increase of the phase path in a planetary atmosphere and ionosphere may be written in a form used by Fjeldbo *et al.* (1965) (using a cylindrical coordinate system, with the zero point in the center of a planet, and the z-axis directed along the earth–planet line – Figure 1 –, and supposing that outside the planetary atmosphere and ionosphere the refraction index is equal to unity)*:

$$\Delta\varphi(\varrho, f) = \frac{1}{\lambda}\int_{-\infty}^{\infty}(\mu_n - 1)\,\mathrm{d}z + \frac{1}{\lambda}\int_{-\infty}^{\infty}(\mu - 1)\,\mathrm{d}z\,, \tag{1}$$

where λ is the wavelength in vacuum, μ_n the refraction index in the neutral atmosphere; μ is the refraction index in the ionosphere which, if one assumes that the frequency f (cycles per second) is much greater than the electron's cyclotron and plasma frequency as well as the collision frequency of the electrons with neutral particles, is equal to

$$1 - \frac{4.03\times 10^7 N_e}{f^2}.$$

It is assumed that in the lower atmosphere $\mu = 1$ and the altitude profile $\mu_n(h)$ is determined, while in the upper atmosphere $\mu_n = 1$ due to the small density of the atmosphere, and the profile $\mu(h)$ and hence the $N_e(h)$-profile is determined. Refraction can be ignored if

$$\frac{4.03\times 10^7 N_e}{f^2} \ll 1\,.$$

The altitude profile of the refraction index can be obtained by inverting the integral Equation (1) either by selecting a profile satisfying the measured increase of the phase path or through an approximate transformation, in which the integral increase of the phase path in the ionosphere is replaced by a sum of increases of the phase path in a finite number of layers of the ionosphere or neutral atmosphere, each of which is characterized by a constant refraction index. In both methods it is supposed that

* The acceptability of the last assumption is illustrated by the experimental data given in Section 2; see Figure 3a.

the ionosphere and atmosphere are spherically symmetric, but the second method allows one to obtain a less smoothed profile of the refraction index and consequently some features of the fine structure of the atmosphere.

In a dual radio-frequency experiment the unambiguity due to the simultaneous influence of the neutral and charged particles on the refraction index is ruled out since the dispersion effect allows one to single out the contribution of the ionosphere to the relevant phase increase. For high frequencies, which are usually chosen as multiples of a low frequency ($=mf$), the increase of the phase path is given as (Fjeldbo *et al.*, 1965).

$$\Delta\varphi(\varrho, mf) = \frac{m}{\lambda}\int_{-\infty}^{\infty}(\mu_n - 1)\,\mathrm{d}z - \frac{4.03\times 10^7}{mf}\int_{-\infty}^{\infty} N_e\,\mathrm{d}z\,. \tag{2}$$

Apparently the system of Equations (1) and (2) with two unknowns permits one in principle to determine both μ_n and N_e.

Such experiments can be performed by two methods, either by the emission of radio waves generated aboard a spacecraft or through the re-emission of radio waves received in the spacecraft from the earth (with possible transformation of their frequency). In the second case the frequency of the radio waves is more stable since it is determined by ground-based frequency standards which increases the accuracy of phase measurements.

2. Experimental Results of Investigations of the Ionospheres of Mars and Venus

In July 1965 an experiment was conducted for observing the radio occultation of the American space probe Mariner 4 by the planet Mars and for the first time the characteristics of the Martian ionosphere were obtained (Kliore *et al.*, 1965; Fjeldbo and Eshleman, 1968). Single frequency (2100 MHz) radio waves were used with re-emission of the radio waves, received from the earth in the spacecraft. Figure 2 shows the total phase variation, measured in the experiment, during the passage of radio waves through the ionosphere (the negative values of $\Delta\varphi$ increase at most to about 10 cycles*) and through the planet's lower atmosphere (the maximum positive value of $\Delta\varphi$ is about 30 cycles). Figures 3a and 3b show the altitude variation of values of the refractivity $N=(\mu-1)\times 10^6$ during immersion (day) and emersion (night) of the spacecraft determined from an inversion of Equation (1) while replacing the integration by a sum over a finite number of layers. Figure 3a shows some examples of the refractivity profiles obtained prior to the radio occultation to illustrate the degree of deviations of the refraction index from unity during the propagation of the radio waves between the earth and the spacecraft before reaching the planetary atmosphere.

Let us consider in more detail how the refractivity was determined from the phase data. As stated, the values of $\Delta\varphi$, measured during the passage of the waves through

* 1 cycle corresponds to 2π.

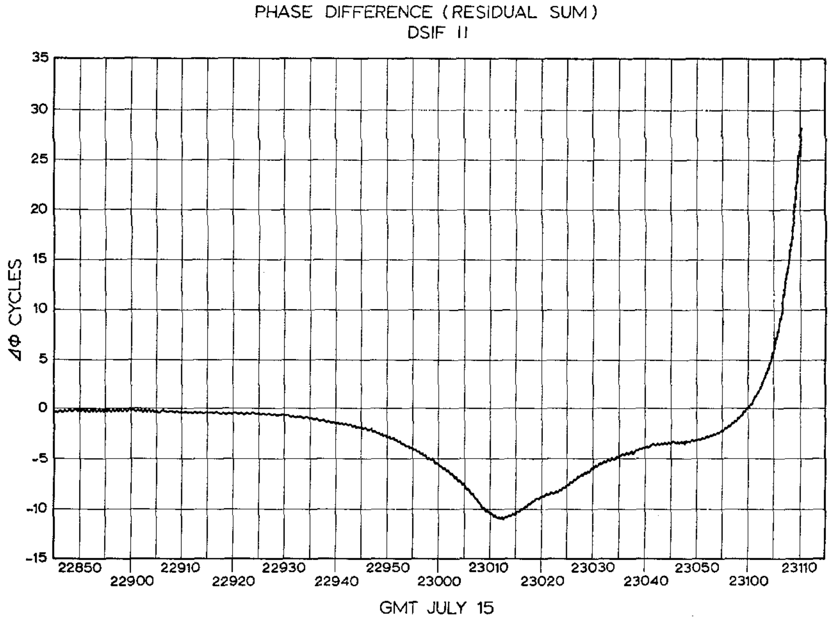

Fig. 2. The total variation of the phase as a function of time, during the pass of radio waves with a frequency of 2100 MHz through the atmosphere of Mars (Kliore *et al.*, 1965).

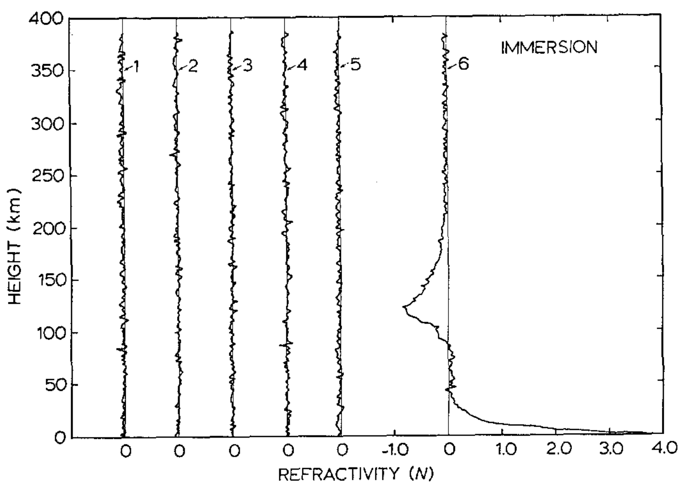

Fig. 3a.

Fig. 3. Refractivity profiles, obtained: (a) before the beginning of the occultation (1, 2, 3, 4, and 5) and during the immersion (6) of Mariner 4 into occultation by the atmosphere of Mars (the ordinate scales of curves 1–5 are conditional); (b) during emersion (on the night-side of the planet) (Fjeldbo and Eshleman, 1968).

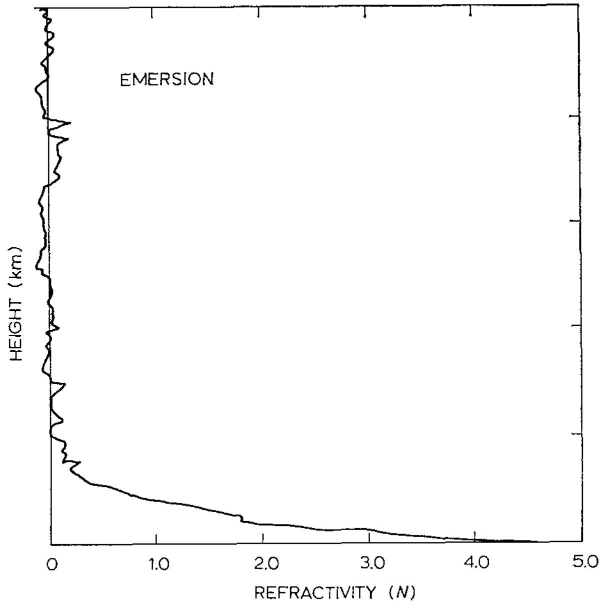

Fig. 3b.

the planet's atmosphere, contain a contribution caused by the geometric increase of the distance of the spacecraft, variations of the earth's ionosphere and the interplanetary medium, as well as the influence of a radio system (the reference signal phase deviations, etc.). The trajectory of the vehicle before its occultation by the atmosphere was calculated from Doppler measurements. Subsequently, when the occultation began, the trajectory was extrapolated with a high accuracy (since the spacecraft moves ballistically) and the increase of the phase path due to the geometric increase of the vehicle's path determined by this extrapolated trajectory was calculated, assuming that the motion takes place in vacuum.

To take into account the influence of variations of the terrestrial ionosphere, the interplanetary medium and the instability of the radio system on the measured increase of the radio-phase during the occultation the following was done. The length of the occultation was ~ 1.5 min (see Figure 2). Variations in the phase of the received radio waves were considered during some of such time intervals before the beginning of the passage of radio waves through the planetary atmosphere (see Figure 3a, curves 1, 2, 3, 4 and 5). Apparently such variations are much less than the effects produced by the Martian atmosphere. Variations corresponding to the time interval directly preceding the occultation were averaged and subtracted from the phase variation measured during the occultation. The fact that the effects of the earth's ionosphere and the interplanetary medium are much lower than the effects produced by the atmosphere of the planet becomes quite clear, if one takes into account that during the occultation the change of direction of the radio beam in the earth's ionosphere is very small (see Figure 1), and that only time variations of the integral electron density in the earth's ionosphere have a practical effect. Over the same time interval the variation of the electron density along the beam, due to the influence of the planetary ionosphere, fluctuated from zero to more than twice the integral electron density in the planet's ionosphere (see Figure 1). It is evident from Figure 3b, that the charged particle density in the Martian night-time ionosphere is below the level of sensitivity of the experiment.

The solid line of Figure 4 shows the altitude profile of the electron density in the Martian ionosphere plotted from data on the profile of the refractivity obtained in the first publication (Kliore *et al.*, 1965) by selecting a model, while the plots in Figure 5 are based on data represented in Figure 3a (Fjeldbo and Eshleman, 1968). Figure 6 shows the altitude-temperature profiles calculated from the scale heights of the electron density profile, shown in Figure 5. The temperature profile was derived assuming, that $T_e = T_{\mathrm{neutr.}}$ and that the value of the average ionic mass and the value of the temperature at the initial level of integration are known.

The probe disappeared behind the limb of the planet at a Martian latitude of 50°, where the solar zenith angle was about 67° and the season corresponded to the end of the winter. On the basis of the results represented in Figures 3, 4 and 5 the following conclusions may be drawn:

(1) The maximum electron density in the daytime ionosphere of Mars at middle latitudes occurs at a height of 125 km and is 9×10^4 to 10^5 cm^{-3}.

(2) The electron density scale height above the maximum can be regarded as approximately constant and equal to 20–25 km.

(3) A small secondary maximum is observed at a height of about 20–25 km below the main maximum.

(4) The electron density in the night-time ionosphere of Mars does not exceed 10^4 cm^{-3}.

The electron density vs. altitude profiles shown in Figures 4 and 5 have been obtained by a single-frequency method, i.e., as mentioned above, by a method which does not allow one to unambiguously separate in the lower atmosphere the contributions to the value of the refraction index of neutral and charged components. Harrington *et al.* (1968), in particular, showed, that if near the surface of the planet (at a height of about 15 km) there is a region with a charged particle density, which is the same or by an order of magnitude smaller than that one in the detected maximum of the ionosphere (see Figure 4), this may considerably change the resulting scale height of the lower portion of the Martian atmosphere (varying within the limits of 9–15 km), and yet satisfying the measured altitude profile of the refraction index. In Figure 7 the electron density profile for a scale height $H_{\mathrm{neutr.}} = 9$ km is indicated by curve 1 and for $H_{\mathrm{neutr.}} = 15$ km by curve 4, respectively.

It is probable that there is an enhanced electron density region near the surface of Mars, since Mars has no intrinsic magnetic field and consequently there is a possibility, that solar wind particles (apparently disturbed by interaction with the planetary ionosphere) can reach sufficiently dense layers of the atmosphere and ionize them. The second cause may be the absence of large amounts of O_3 and O_2 in the Martian

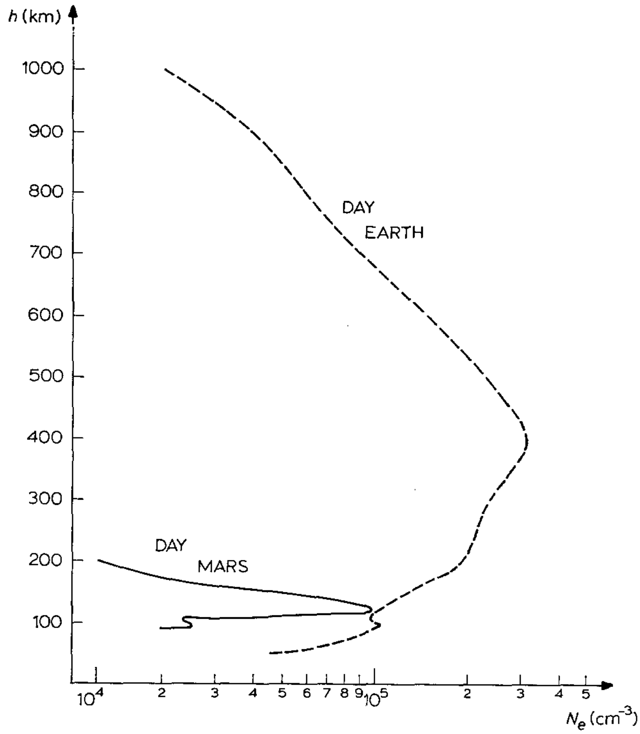

Fig. 4. Altitude distributions of the electron density in the ionosphere of Mars (solid curve) and the earth (dotted line). The density profile in the Martian ionosphere is obtained from data given in Figure 2 (Kliore *et al.*, 1965).

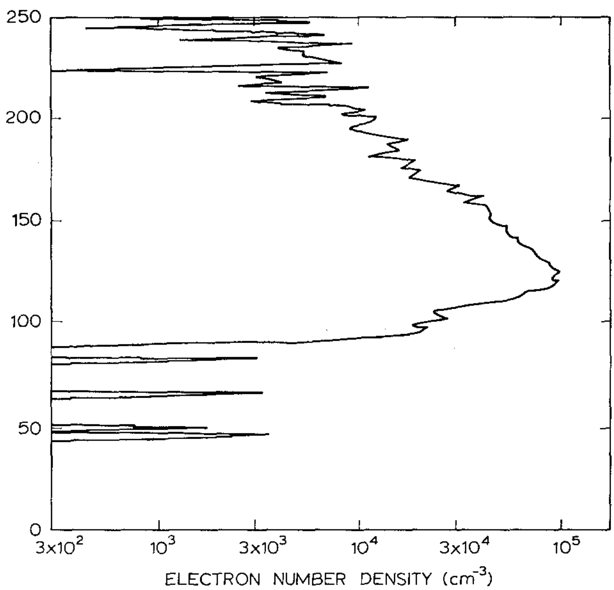

Fig. 5. Electron density profile for the Martian ionosphere, obtained from data, given in Figure 3 (Fjeldbo and Eshleman, 1968).

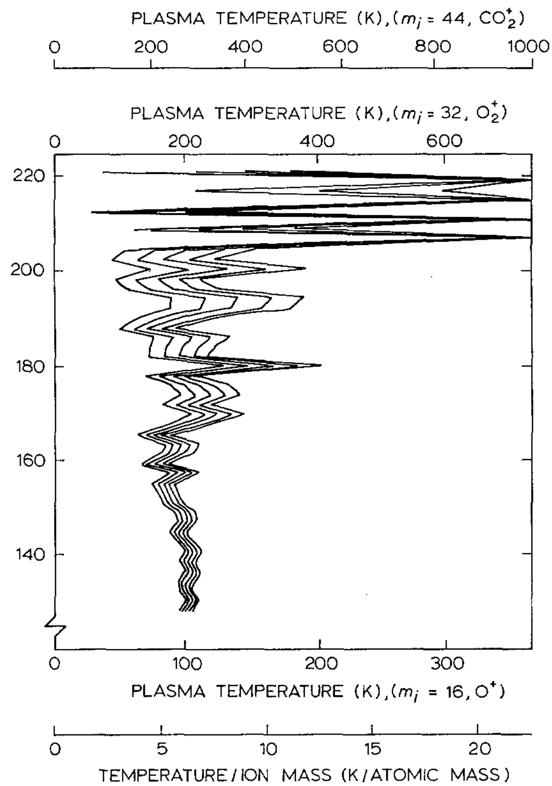

Fig. 6. Altitude temperature profiles for the Martian atmosphere calculated from the scale height of the electron density profile shown in Figure 5 for various values of the temperature at the initial level of integration (Fjeldbo and Eshleman, 1968).

atmosphere, the former of which would be able to absorb solar short-wave emission and the latter has a high affinity to electrons and would be able to decrease the free electron density in the lower atmosphere due to their attachment.

Finally it must be noted that the electron density profiles have been obtained on the assumption of spherical symmetry which may be absent in the real atmosphere. All the above-mentioned limitations of the method lead to the fact that the electron and neutral particles density profiles obtained in this experiment should be regarded as an averaged picture of the particle distributions in the Martian ionosphere and it is probable that measurements *in situ* will give results which will differ in some degree from those obtained.

Let us now consider the characteristics of the ionosphere of Venus, determined from the radio-occultation experiment of the American vehicle Mariner 5 on October 19, 1967 (Mariner Stanford Group, 1967; Kliore *et al.*, 1967) and from measurements by means of charged particle traps installed on the Soviet space probe Venera 4 which conducted measurements *in situ* in the planetary atmosphere on October 18, 1967 (Gringauz *et al.*, 1968).

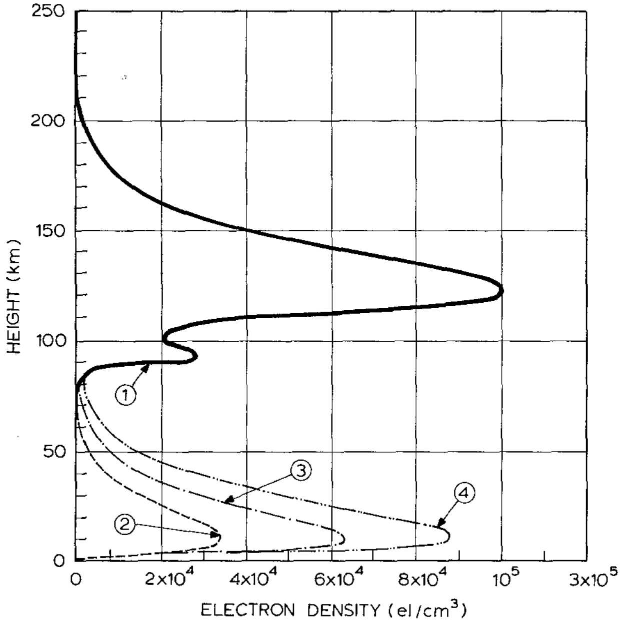

Fig. 7. Plausible electron density profiles in the Martian ionosphere corresponding to the following scale heights of the atmosphere; (1) to 9 km; (2) to 11 km; (3) to 13 km and (4) to 15 km. (Harrington *et al.*, 1968.)

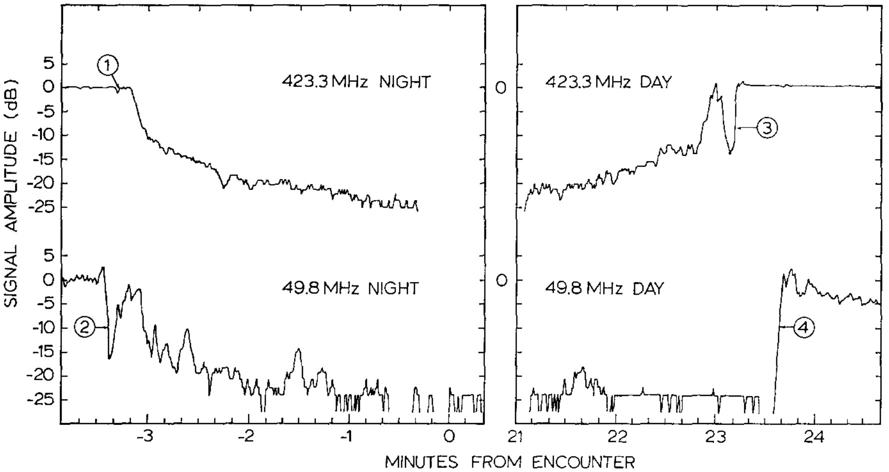

Fig. 8. Signal amplitudes at two frequencies (49.8 MHz and 423.3 MHz) during entry (night-side) and exit (day-side) occultation of Mariner 5 by the atmosphere of Venus. The ionospheric effects are marked 1, 2, 3, and 4 (Mariner Stanford Group, 1967).

In one of the experiments with Mariner 5 carried out by the Stanford University group (U.S.A.), radio waves of two frequencies (49.8 MHz and 423.3 MHz) were used (Mariner Stanford Group, 1967). As stated the dual-frequency experiment in principle permits one to unambiguously determine μ_n and N_e. However, the authors of the experiment point out that the single-frequency experiment with radio waves with a frequency of 2298 MHz (Kliore *et al.*, 1967), which was conducted simultaneously permitted one to obtain better results for the lower atmosphere.

Figures 8a and 8b show dual-frequency variations of the amplitude of the signal during the occultation of Mariner 5 by the planetary atmosphere. The time (minutes) is calculated from the moment of closest approach of the spacecraft to the planet. The moments of passage through the ionosphere are indicated by the arrows 1 and 2, 3 and 4. The dense neutral atmosphere caused the overall slow decrease of the amplitude. Figure 9 shows the profile of electron density vs. height in the daytime and nocturnal Venusian ionosphere obtained by the Stanford group. The dashed curve in the daytime profile indicates its sections plotted on the basis of approximate calculations from recordings of the amplitudes with the use of the model. The dotted curve represents an extrapolation in that region of height where due to the formation of caustics there was no communication. The latitude at which the occultation begins was 37° and the solar zenith angle was 142°. The latitude of emersion was 32°S and the solar zenith angle there was 33°.

The altitude electron density profile in the day-time ionosphere, like the refraction index in Venus's troposphere, was determined from data on the above-mentioned experiment at a frequency of 2298 MHz (Kliore *et al.*, 1967). The instrumentation appeared not sensitive enough to measure the night-time electron densities in the

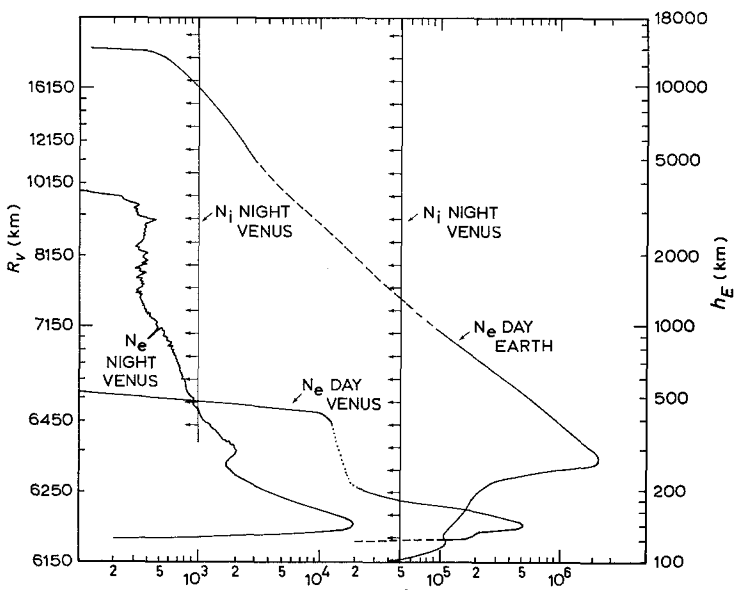

Fig. 9. Altitude profiles of the electron number density in the day- and night-ionosphere of Venus, obtained from data of the Mariner 5 dual-frequency experiment; the upper limits of the ion density in the atmosphere of Venus from data obtained by the Venera 4 probe and the altitude profile of the electron density in the earth's ionosphere (see the text for explanation). (Breus and Gringauz, 1968.)

ionosphere of Venus. Unlike the similar experiment in Mariner 4 in this case the single passage of a signal from the spacecraft to a ground-based station was used. In Figure 10 the profile of the electron density vs. height in the day-time ionosphere of Venus, as obtained in this experiment is shown. In Figures 9 and 10 the distances from the planet's centre are shown.

In Figure 9 straight lines with arrows indicate the upper limits of ionic density in the Venusian nocturnal ionosphere determined from measurements by charged particle traps in the space probe Venera 4.

Planar and hemispherical three-electrode charged particle traps were installed in Venera 4 (Gringauz *et al.*, 1968; Breus and Gringauz, 1968).

The planar traps were designed for measuring low densities in the outermost part of the ionosphere of Venus. Measurements were performed once every 7 sec. To differentiate between ionospheric ions and ions of non-ionospheric origin (e.g., the solar wind) once every 14 sec a positive voltage of +50 V, which would retard thermal ionospheric ions, was applied to one of the grids of the planar trap. At the vehicle's speed of about 10^6 cm sec^{-1} in the section of the trajectory near the planet, the planar traps permitted one to take one measurement per 150 km. The sensitivity limit of the planar traps was 50–5000 cm^{-3}.

Hemispherical traps were designed for measuring positive ions near the main ionization maximum and their sensitivity was chosen within 10^4–10^7 cm^{-3}, in accordance with the hypotheses mentioned in the beginning of this review. To obtain a more detailed altitude ion density profile it was decided to take measurements once every 0.8 sec.

The measurements performed by the hemispherical traps showed that the ion

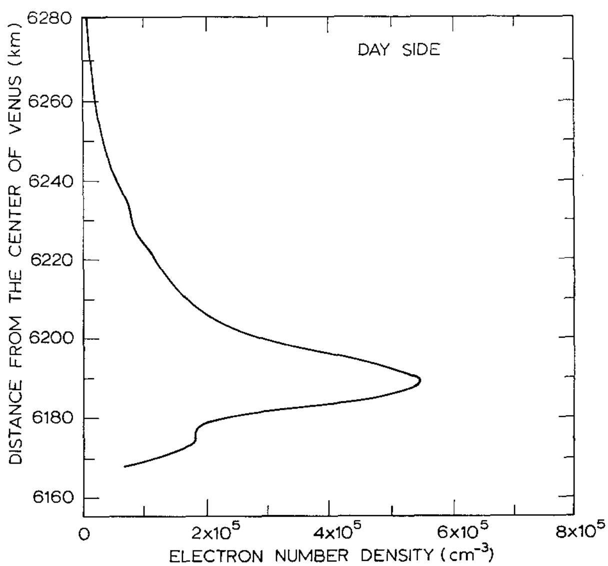

Fig. 10. Electron density in Venus's day-time ionosphere, obtained from data of Mariner 5 single-frequency experiment (S-band). (Kliore *et al.*, 1967.)

density in the night-time ionosphere nowhere exceeded 5×10^4 cm^{-3}. Measurements in the low ion density region were made with 150 km intervals and permitted one to establish an upper limit of about 10^3 cm^{-3} at heights above 300 km. Data from the two last planar trap measurements at lower altitudes could not be decoded due to radio interference.

According to data obtained from Mariner 5 the electron density in the nocturnal ionosphere at heights lower than 300 km (Figure 9) exceeds 10^3 cm^{-3} only in a thin layer of about 150 km and consequently could remain unnoticed during measurement from Venera 4. As is evident from Figure 9 the results of measurements of N_e and N_i in the Venusian night-time ionosphere do not contradict each other.

In the day-time ionosphere the electron density maximum is (5–6) $\times 10^5$ cm^{-3} near 6190 km (about 140 km). There is an additional maximum 15 km below the main maximum. The scale height increases with height which means either an increase of temperature, or a decrease of the ionic mass when light ions start to predominate at greater heights and perhaps both phenomena play a part at the same time.

3. A Comparative Discussion of the Ionospheres of the Terrestrial Group of Planets

Before discussing models of the charged and neutral components of the Martian and Venusian atmospheres as given by some authors on the basis of results of the experiments described above, we shall briefly outline the main characteristics of the earth's ionosphere of importance for a further comparison.

It is well known that the earth's ionosphere extends rather far: its outer boundary lies at distances of about 15 000–30 000 km. However, charged particles are formed in it mainly below 300 km and ascend to greater altitudes due to diffusion along the magnetic lines of force. The magnetic field has a great influence on the distribution of charged particles above the ionization maximum and in defining various regions of the earth's ionosphere. A number of peculiarities of the global distribution of ionospheric charged particles and processes affecting the ionospheric physical properties (e.g., the precipitation of energetic charged particles from the radiation belts and the magnetospheric tail into the lower ionosphere causing its heating and additional ionization) are associated with the geomagnetic field. Not only the transportation of charged particles takes place along the magnetic field tubes, but also the transfer of thermal energy from the sunlit part of the earth's ionosphere into the nocturnal part.

The main layers with enhanced electron density in the earth's lower ionosphere, called the F_2, F_1, E and D-regions, have different origins and are characterized by specific features in the time-dependent and spatial variations of their parameters.

The main maximum in the terrestrial ionosphere in F_2 region is formed above the level of the maximum rate of ion formation, due to the fact that the ion-molecular reactions and the subsequent dissociative recombination of the formed molecular ions are the main causes of a reduction of the number of charged particles in the earth's ionosphere below 200 km. The concentration of atmospheric molecular components (and hence the effective rate of recombination) decreases rapidly with height. The rate of ion formation is proportional to the concentration of the atmospheric atomic component and, therefore, decreases slower with height than the recombination rate. Under such conditions the maximum of the electron density may appear above the region of maximum ion formation; the equilibrium being maintained by the downward transfer of the ion excess formed there to the region of rapid recombination through ambipolar diffusion. The atomic oxygen ion is the most important ion in the F_2 maximum.

Chapman's simple layer theory gives a good approximation for the F_1, E and D regions, the distribution of charged particles is being described by photochemical equilibrium, and the height of the electron density maximum coincides with the height of the ionization rate maximum. The observed division of the ionosphere below the main maximum in two or three regions (F, E and D) with maximum densities increasing from D to F_1 is caused by a combination of the height variations of density and chemical composition of the upper atmosphere and of the spectral intensity distribution of the solar ionizing radiation. E.g., the F_1 region is due to ionization of the atmosphere by ultraviolet radiation with wavelengths shorter than 1027 Å and up to tens of Å. The number of neutral particles in an atmospheric column with a cross-section of 1 cm^2, above the height of the F_1 maximum is approximately 10^{17} cm^{-2}.

Naturally, attempts were made to explain the structure of planetary ionospheres in a similar way as the structure of the earth's ionosphere, and precisely from this standpoint some authors have discussed the experimental results. Difficulties met in these cases are due to the lack of data on the chemical composition and properties of the neutral atmospheres out of which the ionospheres are formed. As pointed out above the determinations of the refraction index from the observations of radio occultations permit the evaluation of the neutral particle density only in the lower atmosphere with the same degree of accuracy with which it is possible to determine the chemical composition from data of other, in particular spectroscopic observations, while the pressure limits near the planetary surface and the main atmospheric composition vary between extremely wide limits according to the estimates of different authors. E.g., the narrowest limits for the CO_2 content in the Martian atmosphere are 80 and 100% (Fjeldbo *et al.*, 1966; Moroz, 1967). The composition, temperature, and concentration of the upper atmosphere are known mainly from theoretical estimates in which data from indirect measurements are used and from the extrapolation of the data referring to low altitudes. Some parameters of the Venusian atmosphere, e.g., such as the scale height of the neutral atmosphere H, its logarithmic gradient ($d \ln H/dh$) and pressure are determined from observations of the occultation of Regulus by Venus for altitudes corresponding to the level of Regulus' occultation (i.e. 120 ± 10 km). However, these data cannot be unambiguously interpreted if the composition or temperature of the atmosphere are unknown. Since there are no data about the planet's neutral atmospheres in the region of ionospheric heights one has to construct models of neutral atmospheres, which on different assumptions, fit to electron density distributions obtained experimentally. The assumption of photochemical equilibrium in the atmosphere makes it possible to describe the electron density profile by models of the type of the terrestrial D, E and F_1 layers. If we assume diffusion to take part in the formation of the ionization maximum, then the ionization maximum in the planetary ionospheres will be similar to the earth's F_2 layer.

Let us next consider the existing models of Martian and Venusian ionospheres in more detail.

If we assume that the main ionization maximum in the Martian ionosphere is formed like the earth's F_2 layer (Fjeldbo *et al.*, 1966; Johnson, 1965, 1966) the Martian upper atmosphere should consist of an atomic component, e.g. O, and have a low temperature (see Figure 11b). Atomic oxygen may be formed in the Martian upper

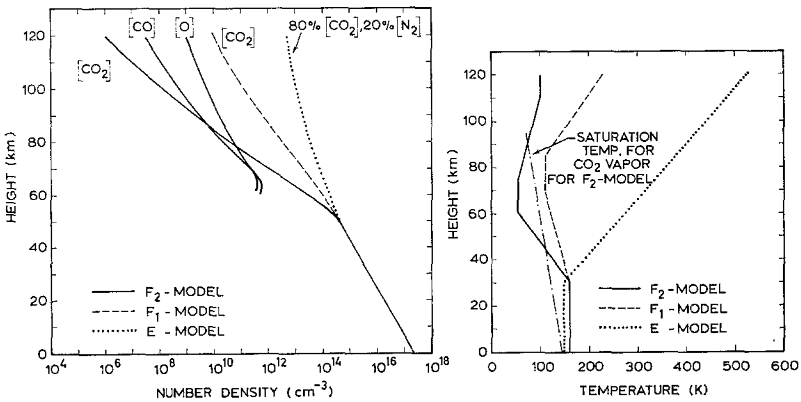

Fig. 11. Models of the Martian neutral atmosphere, calculated from experimental data obtained by Mariner 4, assuming that the maximum electron density is described by a model of the ionosphere of the type of the earth's F_2 and F_1 regions and on the basis of an analysis of the thermal balance in the neutral atmosphere (model E). (Fjeldbo and Eshleman, 1968.)

atmosphere due to dissociation of the main component (CO_2) under the effect of solar radiation at a wavelength of 1700 Å, which takes place somewhere in the height range of 70–90 km. A precise determination of the level of dissociation of CO_2 and the n(O)-profile depends on the knowledge of the optical thickness of the atmosphere at a height of about 100 km and, hence, is impossible without a knowledge of the profile of the total density of the neutral particles, or of $n(CO_2)$.

The temperature of the neutral atmosphere at the altitude of the maximum should be about 80° to satisfy the measured scale height of about 25 km, when the mean ion mass is 16 a.e.m. (which corresponds to O^+). One assumes that in the Martian ionosphere $T_e = T_i = T_{\mathrm{neutr.}}$ since the ionization occurs at a low altitude where the collision frequency of electrons with the molecular components is sufficiently great. Between these regions the temperature appears to be close to the freezing temperature of CO_2. This helps in solving the problem arising due to the high reaction rate of $O^+ + CO_2 \rightarrow O_2^+ + CO$, which may be the main reaction responsible for the disappearance of O^+ ions in the ionosphere of Mars. At the rate of this reaction of about

10^9 cm^2 sec^{-1} (Fehsenfeld *et al.*, 1966) O^+ ions would rapidly turn into O_2^+ ions and the F_2 maximum, with a density exceeding that of F_1, might not be formed. But if a process like CO_2 freezing of the temperature minimum near 100 km would decrease the density of CO_2 molecules, the contradiction would be solved.

The drawbacks of the model of the Martian ionosphere similar to the terrestrial F_2 layer are the following. From theoretical calculations of the thermal balance in the Martian upper atmosphere (Chamberlain and McElroy, 1966) it follows that on Mars a region with a temperature increase should exist with a temperature of 400 K at a height of about 300 km (the thermosphere). Such a region cannot exist if the ionosphere of Mars is described by the model of the F_2 type. Further, the Venera 4, 5 and 6 experiments have shown that although CO_2, as on Mars, is the main component in the lower atmosphere of Venus, its dissociation into CO and O is apparently not essential since in the Venusian upper atmosphere O atoms have practically not been detected.* If the processes in the upper atmosphere of Mars are similar to those on Venus this fact is not in favour of the F_2 model.

In Figures 11a and 11b the distributions of components of the Martian neutral upper atmosphere and its temperature are shown which allows one to build ionospheric models of the type of the terrestrial F_2, F_1 and E layers (Fjeldbo and Eshleman, 1968). The figures show that the nocturnal particle densities in these three models at a height of 120 km differ by four orders of magnitude. The temperature also differs sufficiently especially for the F_2 and E models.

In the model of the Martian ionosphere of the type of the terrestrial E region high temperature values and a considerable positive temperature gradient in the upper atmosphere occur, which corresponds to theoretical estimates (Chamberlain and McElroy, 1966; McElroy, 1969). However, this model involves even more contradictions than the F_2 model. E.g., to explain the absence of the F_1 and F_2 layers above the E region, one has to suppose that the upper atmosphere is mixed to a height of 190 km and that the coefficient of the effective ionic recombination rate increases with height. This contradicts the situation observed in the earth's ionosphere. For appreciable mixing of the atmosphere to such heights there must be a considerable downward heat flow, which rules out the high temperatures in the upper atmosphere, obtained in the preceding model from thermal balance without taking turbulent heat transfer into account. Finally, another point of contradiction between the model and experimental data is the fact that the experimental electron density profile, as pointed out above, has a constant scale height above the maximum, from which it follows that there is no region with an increasing temperature above the maximum (up to the height of about 250 km), or that a slow rise in the temperature takes place which is compensated for by the decrease of the average ionic mass.

In the E model (Chamberlain and McElroy, 1966) (for a mean molecular mass of

* The molecular oxygen content in the lower atmosphere of Venus is less than 1 % and in the upper atmosphere at altitudes of $\gtrsim 400$ km the instrument, designed for measuring the glow of the Venusian atmosphere in the resonance oxygen lines, did not detect this glow (Vinogradov *et al.*, 1968; Moroz and Kurt, 1968; and *Izvestiya Verkhovnogo Sovieta SSSR*, No. 130 (16135), June, 4, 1969.

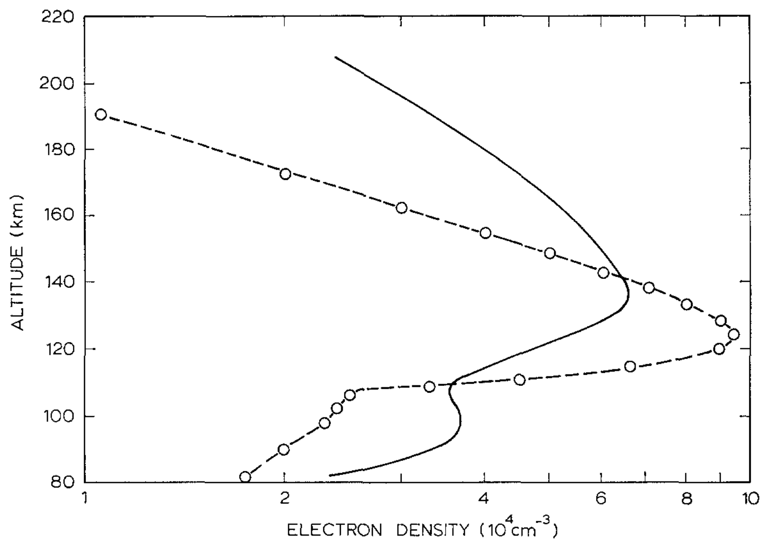

Fig. 12. Experimental profile of the electron density and the profile calculated for a model of the Martian neutral atmosphere, obtained from an analysis of its thermal balance (McElroy, 1969).

35 a.e.m.) a rise in the temperature takes place in this height range by more than a factor of 2. Figure 12 shows (solid curve) the altitude profile of the electron density in the ionosphere of Mars (McElroy, 1969) calculated for a model of the neutral atmosphere consisting of pure CO_2 obtained by McElroy (1969) by a method similar to that used by Chamberlain and McElroy (1966). Apparently, the scale height of this profile, equal to 58 km, differs from the experimental scale height of the electron density (dotted curve), by more than a factor of 2. If the average ionic mass becomes lower than 44 a.e.m. due to an increasing number density of ions lighter than CO_2^+, the difference will be still greater.

At last, there is a model of the type of the terrestrial F_1 layer (Donahue, 1966). As this model is based on the distribution of the neutral particle density and temperature obtained in the E layer type model (Chamberlain and McElroy, 1966), but with a mean molecular mass not of 35 (44% of CO_2^+ and 45% of N_2^+), but of 44 a.e.m. (i.e. corresponding to 100% of CO_2^+), temperature in the ionosphere will be $\gtrsim 235$ K (see Figure 11).

Evidently the final choice of the model is difficult at present. It presupposes knowledge of the temperature of the neutral particles and the composition of the upper atmosphere or, at least, of one of these characteristics.

For Venus the interpretation of the results of measurements of the electron density profile in the daytime ionosphere is easier because the Venera 4, 5, and 6 experiments, as pointed out above, have not shown the presence of any considerable concentrations of oxygen atoms in the upper atmosphere. This gives grounds to believe that the ionization maximum obeys the laws of photochemical equilibrium, i.e. may be described by F_1 or E layer type models.

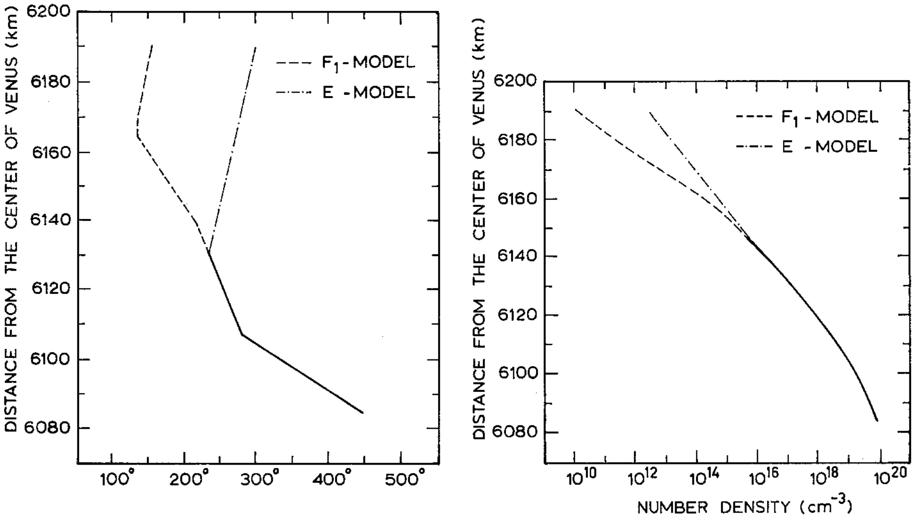

Fig. 13. Models of Venus's day-time neutral atmosphere, corresponding to the models of the ionosphere of the type of the earth's F_1 and E layers (Kliore *et al.*, 1967).

In Figure 13a and 13b, models of the neutral atmosphere satisfying the ionospheric models of the F_1 and E types are shown. At distances less than 6130 km from the planet's centre they were obtained from the refraction index in the lower atmosphere, determined from observation of the Mariner 5 occultation, assuming that it consists of CO_2. The dotted and dashed curves represent extrapolations of these data for models of the ionospheres of the F_1 and E type.

The night-time ionization maximum on Venus may be similar to the D and E layers in the earth's night-time ionosphere. Apparently it is possible to consider two types of ionization sources which can produce such a maximum (McElroy and Strobel, 1969): solar radiation, if the recombination of the day-time ionosphere is sufficiently slow, and nocturnal ionization sources. The first source cannot be effective, since the recombination time at the height of the ionization maximum is of the order of 100–200 sec. Other ionization sources, as cosmic rays, meteorites and Lyman-α radiation diffused by neutral hydrogen, would produce the maximum either at lower, or at higher altitudes than the observed one. E.g., in order to have the ionization maximum similar to the E-maximum in the earth's ionosphere and produced by the ionization of the atmosphere by the above-mentioned sources, the neutral particle density at the height of the maximum should have been approximately equal to 10^{13}–10^{15} cm^{-3}. From the models of the neutral nocturnal atmosphere, constructed by extrapolation of data, obtained in the lower atmosphere, e.g., by Venera 4 (Moroz and Kurt, 1968) it follows that the density at the height of the maximum is much lower (see Figure 16). Charged particle fluxes of about 2×10^7 cm^{-2} sec^{-1} and with energies of about 100 eV for electrons and 1 keV for protons may be one of the ionization sources. A flux of that magnitude corresponds to about 10% of the undisturbed flux of the solar wind, but it is difficult to imagine, how the solar wind flux can get into the planet's night-side lower atmosphere, far from the terminator.

In addition to the causes of the night-time maximum formation discussed above one may suppose also that the nocturnal ionosphere is formed by transfer of charged particles from the planetary day-side to the night-side by winds due to the horizontal temperature and pressure gradients, respectively (McElroy and Strobel, 1969). In that case the neutral particle density at the height of the maximum could be lower than 10^{13}–10^{15} cm^{-3}. In order to retain an electron density of about 10^4 cm^{-3} in the region close to the maximum, where the recombination time is about 200 sec, the particle transfer from the planet's day-side should occur at supersonic velocities (McElroy and Strobel, 1969). The horizontal particle transfer has high speeds in the upper less dense atmosphere and can there contribute considerably to its ionization, but in the region of the ionospheric night-time maximum this source is insufficient and for the explanation of this phenomenon other ionization sources should be involved, such as charged particle fluxes.

There is an interesting peculiarity in the altitude profiles of Venus's electron density: in the night-time profile the electron density remains constant and equal to about 300 cm^{-3} in the distance range of 7500–9500 km and forms a 'knee' (a name that we may suggest for a sharp bend, in analogy with the earth's ionosphere) at distances of 9700 km (see Figure 9). In the day-time profile of the electron density the 'knee' has been detected at distances of about 6570 km and above it the thermal plasma has not been observed. If one supposes that the horizontal transfer of charged particles from the day-side to the night-side maintains the upper nocturnal ionosphere, it is evident, that the effect of this source could not have extended to altitudes of 3000 km, where on the day-side there is no thermal plasma. Perhaps the exact determination of the phase path change produced, e.g., by the deviation of the form of the ionosphere from the spherically symmetric one, would change the N_e-profile to such an extent, that it could have been described by the distribution of H^+ ions in diffusion equilibrium at T of about 1000 K. Since the difference between the maximum N_e values in the day-time and nocturnal ionospheres is approximately of an order of magnitude, the absence of spherical symmetry is evident, while the excessive charged particle content in the outlying regions of Venus's nocturnal ionosphere may be due to the effect of an additional source of ionization, such as e.g. charged particle fluxes (Fjeldbo and Eshleman, 1969). Such an explanation is more suitable for the outer region of Venus's nocturnal ionosphere, than for its region of maximum N_e because this outer region is close to the zone of solar plasma disturbance, produced by a flow of the solar wind around the planet (Bridge *et al.*, 1967; Gringauz *et al.*, 1968), which may be the source of charged particle fluxes ionizing the nocturnal atmosphere at these altitudes (Fjeldbo and Eshleman, 1969).

In the altitude range of 6800–7500 km the shape of the nocturnal N_e-profile may be described approximately by the H^+ distribution at the temperature $(T_e + T_i)/2 =$ 626–1100 K. In the region of 6800–6300 km the most suitable ion can be He^+ at temperatures of 620–970 K and at lower altitudes heavier ions may occur (Mariner

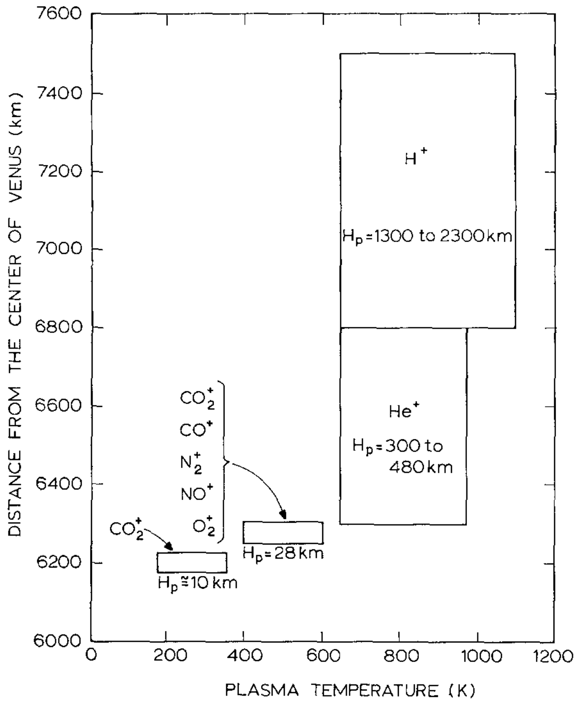

Fig. 14. Possible constituents and temperature for Venus's nocturnal ionosphere, derived from measured scale heights (upper three boxes) and from a preliminary model-fitting to the amplitude measurements (lower box) (Mariner Stanford Group, 1967).

Stanford Group, 1967; McElroy and Strobel, 1969). In Figure 14 the approximate distribution of the composition of ions and the scale height of the electron density profile in Venus's nocturnal ionosphere are in accordance with the estimates, described above (Mariner Stanford Group, 1967). The models shown there of the night-time upper ionosphere from 6400 to 7500 km, consisting of He^+ and H^+ ions or of H^+ and H_2^+, which are in diffusion equilibrium agree well with the experimental profile of the electron density. In Figure 15 one of such models is shown (McElroy and Strobel, 1969). He^+ ions may be formed by photo-ionization of the atoms in the day-time atmosphere and are transferred to the night-side by winds, as noted earlier. The helium content in Venus's atmosphere should by far exceed its content in the earth's atmosphere, which is quite probable even at a rate of formation of He comparable with the terrestrial one. Due to the absence of the magnetic field, Venus has no source of helium loss as the assumed helium ion loss mechanism on the earth – the 'polar wind', i.e. the escape of light ions through unclosed field lines of the magnetospheric tail to interplanetary space (Axford, 1968).

As for the Martian ionosphere, the final elucidation of the regular features of the formation of the Venus ionosphere should apparently be postponed till more reliable data are available about the planet's neutral upper atmosphere.

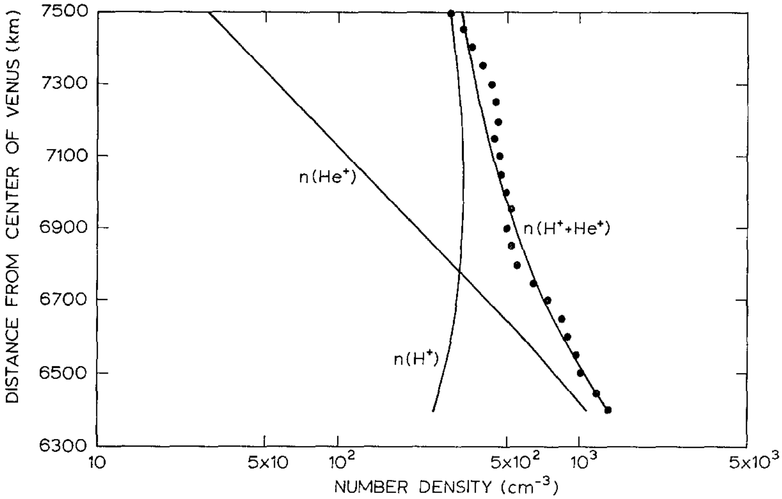

Fig. 15. Model of Venus's night-time ionosphere, consisting of H^+ and He^+ ions diffusive equilibrium, which fits to the nocturnal profile of the electron density, shown by the dotted curve (a portion of the profile shown in Figure 9). (McElroy and Strobel, 1969.)

We next compare the day-time altitude profiles of the electron density in the ionospheres of Mars and Venus with the day-time profiles of the electron density of the earth's ionosphere, corresponding to the same phase of solar activity. To construct profiles of the Earth's ionosphere (see Figure 4, dotted lines, and Figure 9) use is made in the lower ionosphere (up to the main maximum) of results of ground-based sounding, performed on the days of observation of the planetary ionospheres (Ionospheric Data, 1965, 1967). In the outlying ionosphere at heights of 14 000–20 000 km use is made of the measurements of the ion density by means of charged particle traps conducted on satellites of the Electron type in 1964 (Bezrukikh, 1969).

From Figures 3 and 9 it is evident, that the earth has a denser and more extended ionosphere than Mars and Venus, although it occupies an intermediate position between these planets, as far as the distance from the sun and hence the intensity of the ionizing radiation are concerned. The heights of N_e-maxima in the planetary ionospheres lie much lower than the height of the main ionization maximum in the earth's ionosphere. It is worthwhile noting that the maximum day-time and night-time temperatures in the earth's ionosphere are much higher than the corresponding temperatures in Venus's ionosphere and by an order of magnitude higher than the temperature in the Martian day-time ionosphere.

The main reason for these differences is that the densities, chemical composition and temperatures of the neutral atmospheres of the planets around which ionospheres are formed, greatly differ from each other.

In Figure 16 the available experimental and model altitude distributions of the neutral and charged particle densities are given for illustration. In Figure 17 the temperatures in the atmospheres of Mars and Venus are shown. For the sake of com-

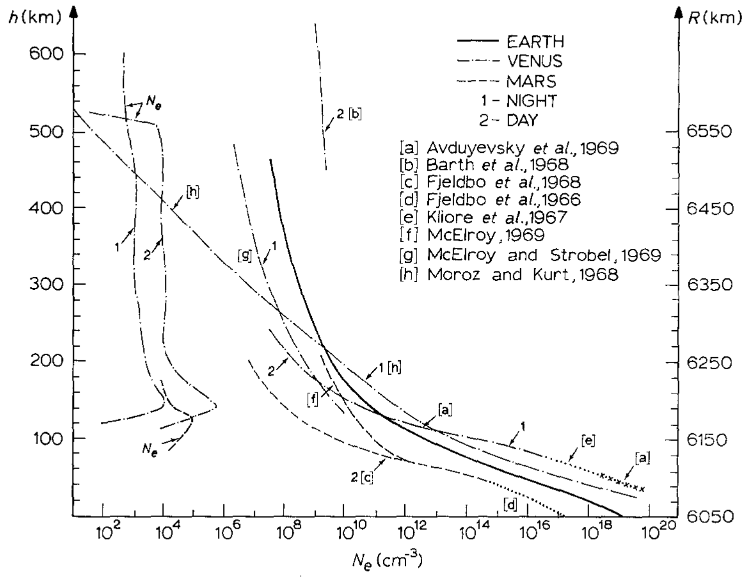

Fig. 16. Model and experimental height distributions of the neutral and charged particle number densities in the atmospheres of Mars and Venus. The values of the neutral particle number densities, obtained from measurements by Venera 4, are indicated by crosses, from measurements by Mariner 4 and Mariner 5 by dots. For the sake of comparison the neutral particle densities in the earth's atmosphere are also given.

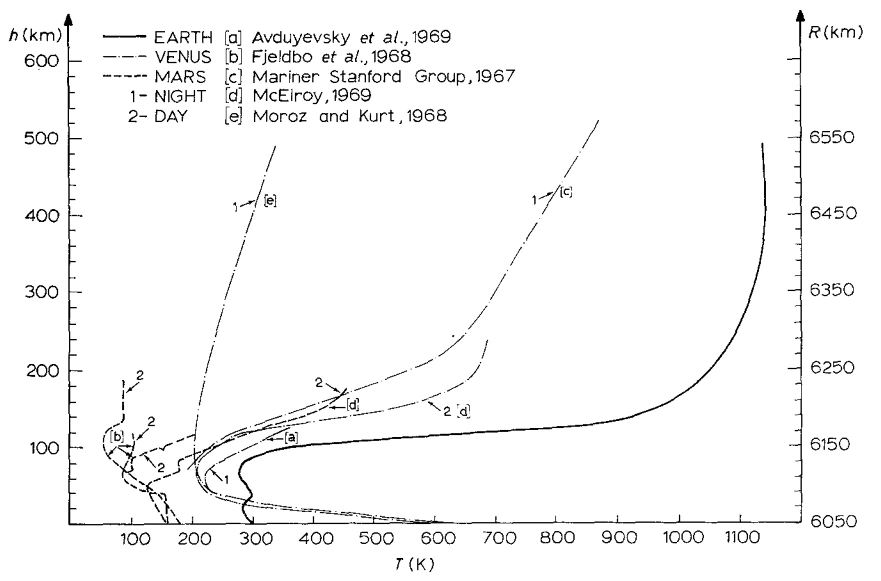

Fig. 17. Altitude distributions of temperatures for the atmospheres of Mars, Venus and the earth from data of different models and from the interpretation of experimental data.

parison the temperatures and densities of the neutral particles in the earth's atmosphere are also given. Models of the upper atmosphere of Mars have been calculated in accordance with the electron density profile obtained in the Mariner 4 experiment (Fjeldbo and Eshleman, 1968). For the night-time upper atmosphere of Venus there is a model of the atmosphere between heights of 100–500 km consisting of atomic oxygen (Moroz and Kurt, 1968), constructed on the basis of the Venera 4 data (Mikhnevich and Sokolov, 1969; Avduyevsky *et al.*, 1969) and there are models in which, starting at altitudes of about 350 km, the atmosphere consists of He and H, and at lower altitudes mainly of CO_2. These models are constructed from the Mariner 5 data and are based further on calculations of the thermal balance in the upper atmosphere (McElroy, 1969; McElroy and Strobel, 1969). For the day-time atmosphere of Venus in the height range up to 200 km the models of the atmosphere, consisting of CO_2, are obtained by extrapolating measurements of Mariner 5, Venera 4 and from Regulus's occultation, with due account of the thermal balance in the upper atmosphere (McElroy, 1968, 1969), and at heights above 400 km the estimates of the H and H_2 (main components) contents in the day-time atmosphere are made from measurements of the Lyman-α radiation by Mariner 5 (Barth *et al.*, 1968). It should be mentioned that in the oxygen model (Moroz and Kurt, 1968) the night-time upper atmosphere of Venus is very tenuous and cold and at altitudes of about 500 km its neutral particle density by approximately an order of magnitude is smaller than the charged particle density. Within the framework of such a model it is difficult to explain the observed constancy of the electron density in the nocturnal ionosphere since disturbed solar wind fluxes would meet an insufficient amount of basic material in the oxygen model of the atmosphere for the ionization of the neutral night-time atmosphere at high altitudes.

In comparing the characteristics of the Martian and Venusian atmospheres with the terrestrial one the following conclusions may be drawn:

(1) The atmosphere of Mars is more rarefied at all explored heights (the particle density near the surface corresponds to about 0.1–0.8% of the particle density in the earth's atmosphere (Kliore *et al.*, 1965)) and is colder than the earth's atmosphere ($T_{\text{neutr.}}$ near the surface is about 180 K and in the region of the main ionization maximum $T_{\text{neutr.}}$ does not exceed 400–500 K (Fjeldbo and Eshleman, 1968), but in the earth's ionosphere $T_{\text{neutr.}}$ is about 800–1200 K (e.g. Polyakov *et al.*, 1968)). The atmosphere of Mars consists mainly of heavy CO_2 gas (Fjeldbo *et al.*, 1966).

(2) The atmosphere of Venus in the region of the main ionization maximum seems to be denser (approximately by an order of magnitude) (Avduyevsky *et al.*, 1969; Polyakov *et al.*, 1968), but colder than the earth's atmosphere ($T_{\text{neutr.}}$ is about 300 K (Avduyevsky *et al.*, 1969)) and consists also mainly of CO_2.

Hence, the atmospheres of these two planets have lower scale heights than the earth's atmosphere (in the vicinity of the main ionization maximum on Mars $H_{\text{neutr.}}$ is about 9 km (Fjeldbo *et al.*, 1966), in Venus it is about 6–10 km (Avduyevsky *et al.*, 1969) and on the earth it is about 25–30 km). Therefore, their ionization maxima lie lower than in the earth's atmosphere. The different evolution of the planet's atmospheres due to their different position with respect to the sun which led to their different carbon dioxide and water contents has played a big role in the formation of these ionospheres.

The absence of intrinsic magnetic fields is another feature, and has also affected the ionospheric characteristics of Mars and Venus. As mentioned above, the boundary of the earth's plasma envelope – the 'knee' in the altitude density distribution (see Figure 9) lies at altitude of 15 000–30 000 km, and varies with changes in the magnetic activity. It separates a portion of the magnetospheric plasma, rotating together with the earth, from the plasma performing large-scale convective motions in the outer parts of the magnetosphere (Axford, 1969).

As noted above in the day-time electron density profile of Venus's ionosphere the 'knee' is observed at a distance of about 500 km from the planet's surface (see Figure 9). The planet has no intrinsic magnetic field. Therefore, the role of the solar-wind stopping obstacle is played in this case by the ionosphere, which has a high electrical conductivity and which prevents the solar wind magnetic field from rapidly diffusing towards the planet's surface. Thus, the disturbance of the magnetic field, which is produced here, may be regarded, by analogy with the earth's magnetosphere, as a magnetopause; it lies, as stated above in the height range of only 500–600 km from the surface on the planet's sunlit side.

On the planet's night side the 'knee' region in the altitude profile of the electron density also seems to be adjacent to the zone of solar plasma fluxes, disturbed by the planet. This zone lies at greater distances from the planet.

For the Martian day-time ionosphere, using the Mariner 5 data it was possible to plot the altitude profile of the electron density only to heights of about 250 km. However, since Mars, like Venus, has no intrinsic magnetic field, but has an ionosphere with a charged particle density comparable to the ionosphere of Venus, it may be stated, that the interaction of the planet with the solar wind creates in the vicinity of the planet a situation similar to that on Venus, and the thermal plasma boundary on Mars lies also much lower than on the earth. The above-mentioned peculiarities of the physical processes in the earth's ionosphere due to the existence of the geomagnetic field, affecting the global distribution of the terrestrial ionospheric plasma are apparently absent in the ionospheres of Mars and Venus due to the absence of intrinsic magnetic fields.

4. Conclusions

(1) Investigation by space probes have made it possible to obtain the altitude distributions of the charged particle densities on the sunlit and dark sides of Venus and in the daytime ionosphere of Mars.

(2) The planets' ionospheres appear to be less dense and extended than the earth's and the regions of maximum electron density also lie much lower than in the earth's ionosphere.

(3) The outer boundary of the thermal plasma in the ionosphere of Venus lies much lower than in the earth's ionosphere (at altitudes of about 500–600 km from

the surface on the planet's day-side while in the earth's ionosphere it lies at a height of about 30 000 km), which is mainly due to the fact that Venus has no intrinsic magnetic field.

(4) The earth's magnetic field is responsible for some essential peculiarities of the structure of the earth's ionosphere, which are absent in the ionospheres of Mars and Venus.

(5) A final choice between models of the types of the F_2, F_1 or E layers for describing the Martian and Venusian ionospheres by analogy with the earth's ionosphere is still premature since there are no unambiguous data on the composition and temperatures of the planets' neutral upper atmospheres.

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