

G. S. Ivanov-Kholodnyĭ, A. V. Mikhailov, and N. A. Savich. Formation of the Ionosphere of Mars. People began to construct models of the ionosphere of Mars about ten years ago, initially starting with purely theoretical considerations. When experimental data began to be obtained with space rockets, these models had to be reconsidered radically. However, the models constructed before the experiments of 1971 described well only the region of the electron-concentration maximum at 140 km altitude, which is analogous to the E or F1 region of the Earth's ionosphere. The models of the higher regions of the Martian ionosphere in essence substantially differed from the model of the Earth's ionosphere. For example, in the best-developed dynamic model of Cloutier, McElroy, and Michel,<sup>[1]</sup> the source of ionization is not the shortwave radiation of the Sun, but the solar particle fluxes. Here the ions and electrons of the ionosphere blow off at supersonic velocity toward the surface of the planet. Why did such unusual models of the ionosphere of Mars appear? Basically there are two reasons: too poor an accuracy of the first measurements of the ionosphere in 1964, and too crude notions of the parameters of the neutral atmosphere and the magnetosphere of Mars at high altitudes.

The need to reexamine the problem afresh arose after the experiments of 1971, which were performed with the space stations "Mars-2" and "Mariner-9." They established the following three important facts: 1) Sh. Sh. Dolginov and his associates<sup>[2]</sup> discovered on Mars a magnetic field whose intensity was so great that it could retain an ionospheric plasma just like that of Earth. 2) Exact measurements of the ionosphere were conducted to an altitude  $h \approx 330$  km.<sup>[3]</sup> Here the electron-concentration profile  $n_e(h)$  (the points in Fig. 1) in the altitude region  $\sim 210$  km exhibits a clearly-marked break that divides the ionosphere into two different parts. In the lower part at altitudes of 150–200 km, the altitude scale  $H_1 = 36$  km, while in the region 220–310 km, it has the  $\sim 1.5$  times higher value  $H_1 \approx 57$  km. This effect had been practically undetectable, owing to the poor accuracy of the earlier measurements.<sup>[4]</sup> 3) It was found from the spectrometric observations of Barth

in 1969<sup>[5]</sup> and 1971<sup>[6]</sup> that the temperature of the neutral atmosphere is 325–350°K (i.e., 1.5–2 times lower than had been assumed) and that there is a slight admixture of atomic oxygen of  $\sim 2\%$  at altitudes  $\sim 140$  km in the atmosphere, which consists mainly of CO<sub>2</sub>.

We can directly conclude from these data that the ionosphere of Mars is "trapped" by the magnetic field, and must be constructed according to the terrestrial model. Below  $\sim 210$  km, it is formed by the action of photochemical processes, while diffusion processes become decisive higher up. The fundamental equation that describes the altitude distribution  $n_e$  is written, as usual, as:

$$\frac{\partial n_e}{\partial t} = q - L + D, \quad (1)$$

Here  $q$  and  $L$  are the photochemical rates of formation and disappearance of electrons, while the diffusional term for the case of one predominant ion has the form

$$D = \frac{\partial}{\partial h} \left[ D \left( \frac{\partial n_e}{\partial h} + \frac{n_e}{H_i} + n_e \frac{\partial H_i}{\partial h} \right) \right],$$

Here  $D$  is the diffusion coefficient, and  $H_i$  is the altitude scale for this ion. We can assume that  $q \approx L$  under quasi-steady-state conditions in the "photochemical" region below  $\sim 210$  km.

Of course, the concrete form of the photochemical processes in the Martian ionosphere differs from those of the Earth, although in both cases the source of the primary ions in daytime is the process of photoionization by the shortwave radiation of the Sun. As has now become clear, CO<sub>2</sub><sup>+</sup> ions are mainly produced below  $\sim 200$  km, but mainly O<sup>+</sup> ions above  $\sim 200$  km, and in general form

$$q = I_0 (\sigma_1^i n_1 + \sigma_2^i n_2) \exp(-H_1 \sigma_1 n_1 - H_2 \sigma_2 n_2),$$

Here  $I_0$  is the intensity of solar radiation, the  $n$  are the concentrations, and the  $\sigma$  and  $\sigma^i$  are the cross-sections for absorption and ionization of O atoms (subscript 1) and CO<sub>2</sub> molecules (subscript 2). The primary ions that are formed enter into various ion-molecule reactions whose result amounts to the following. Owing to the admixture of atomic oxygen, the CO<sub>2</sub><sup>+</sup> ions are efficiently

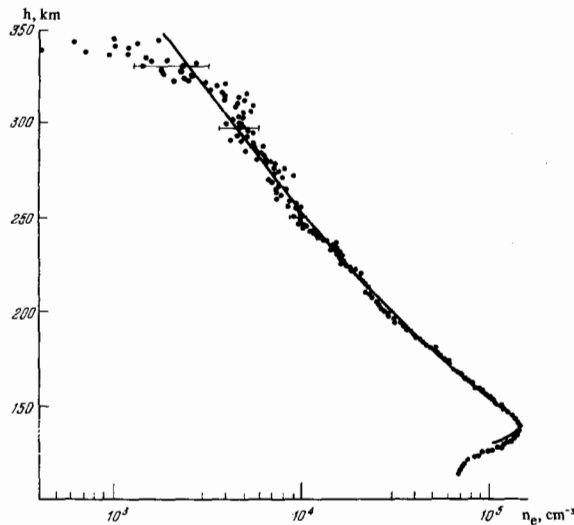


FIG. 1.

converted into  $O_2^+$  and  $O^+$  ions. All of the  $O^+$  ions are converted into  $O_2^+$  ions upon interacting with  $CO_2$ . Hence, the equilibrium ions that must be observed in the ionosphere of Mars are mainly the molecular ions  $O_2^+$  and  $CO_2^+$ , and there are 2–4 times more of the former than of the latter. These ions are governed by recombination processes. Hence,

$$L = \alpha n_e^2,$$

where  $\alpha = (\alpha_2 [CO_2^+]/n_e) + (\alpha_3 [O_2^+]/n_e)$ , and  $\alpha_2$  and  $\alpha_3$  are the recombination coefficients of the ions  $CO_2^+$  and  $O_2^+$ . Here we know that  $\alpha_2$  is almost independent of the temperature, while  $\alpha_3 \propto T^{-1}$ .

A knowledge of the mechanism of formation of the ionosphere in the region of action of photochemical processes permits us to use the ionospheric data obtained by "Mars-2" to calculate the parameters of the neutral atmosphere. In order to do this, we must solve the balance equation  $q = L$  at each altitude for the concentrations  $n_1$  and  $n_2$  of neutral particles. Since there are two unknowns, we need a second equation. This is the generalized barometric distribution law. As the lower boundary condition, we use the data for the maximum of  $n_e$ , which give a reference value for the concentration  $[CO_2] = 2 \times 10^{10} \text{ cm}^{-3}$  at an altitude of 140 km. The data on the break in the  $n_e(h)$  profile permit us rather accurately to determine the concentration of atomic oxygen. Fig. 2 shows the results of calculating a model of the upper atmosphere of Mars for  $h \geq 140 \text{ km}$  for two hypotheses on the temperature-dependence of the recombination constant  $\alpha$ . Actually a certain intermediate case is realized, but the law  $\alpha \propto T^{-1}$  is apparently closer to reality, since  $\alpha \approx \alpha_3$  (the results of the calculations will be published in more detail in the journal *Geomagnetizm i Aéronomiya*).

It is interesting to compare the calculated model with the results of spectrometric observations.<sup>[5,6]</sup> It has been found from these data that  $T \approx 325^\circ\text{K}$  (270–440°K) at an altitude  $h \approx 200 \text{ km}$ , while at altitudes of 135–140 km, the concentration of oxygen atoms is  $\sim 2\%$ , which is close to the model in Fig. 2. The observations in the 160-km region also confirm the general photochemical conclusion that there are few  $CO_2^+$  ions, while the major ions are  $O_2^+$  ions. It was noted in<sup>[6]</sup> that the altitude scale amounts to  $\sim 65 \text{ km}$  for emissions of

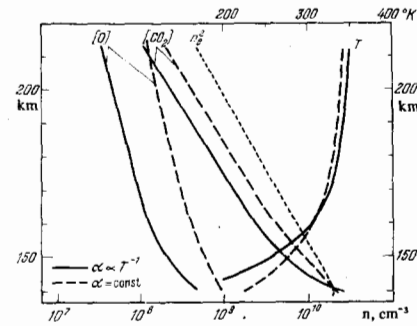


FIG. 2

atomic oxygen above 250 km. This is what should happen, since we can conclude from Fig. 2 that atomic oxygen, whose altitude scale  $H_1 = H_2 \approx 57 \text{ km}$ , predominates in the atmosphere at these altitudes. Thus, the atmosphere model that we have calculated from ionospheric data resembles in many of its parameters that which the spectrometric measurements give.

Let us proceed to analyze the mechanism of formation of the upper part of the ionosphere of Mars. Here we shall try to elucidate why it lacks a layer analogous to the major maximum in  $n_e$  in the F2 region of the Earth's atmosphere. By starting with the neutral-atmosphere model in Fig. 2, which we can easily extrapolate to high altitudes where  $T = \text{const.}$ , we have solved Eq. (1) numerically on a computer (the details will be published in the journal *Kosmicheskie Issledovaniya*). The electron-concentration profile obtained by calculation is shown in Fig. 1 by the solid line. The horizontal bars show the r.m.s. errors of measurement. We see that the theoretical curve agrees well with the results of measurement in reflecting the break in the  $n_e(h)$  profile in the region of  $\sim 210 \text{ km}$ . Thus Eq. (1), which describes the Earth's ionosphere, is also applicable for describing the Martian ionosphere.

Let us now discuss the problem of why Mars lacks a layer analogous to the F2 layer on Earth. As we know, a quadratic recombination law holds in the Earth's ionosphere in the region of action of "photochemistry" in its lower part, while a linear law holds in its upper part. These two laws give a substantial difference in the trend of  $n_e$  with altitude: a decrease in  $n_e$  with increasing  $h$  with a quadratic law, and conversely, an increase with a linear law. This circumstance favors formation of a F2 layer, which arises at a level where the efficiency of the two competing mechanisms is about equal (the photochemical and diffusional mechanisms). Photochemical processes predominate below this level, and hence we observe there an increase in  $n_e$  with increasing  $h$ , while diffusion processes predominate above, which tend to distribute the plasma according to a barometric law. That is, they lead to a decrease in  $n_e$  with altitude. The situation in the Martian atmosphere is such that this possibility is not realized, since the quadratic recombination law  $L \propto n_e^2$  holds in all the regions of photochemical equilibrium. This involves the extremely low concentration of atomic oxygen. If the concentration  $[O]$  were to increase by several orders of magnitude, then the major primary ions would be  $O^+$  ions. In this case, the molecular ions would mainly be formed by ion-molecule reactions, and this would lead to a linear recombination law, and ultimately to formation of a F2 layer. The effect of diffusion is manifested most noticeably in the appearance of the break in the  $n_e(h)$  profile in the region  $\sim 210 \text{ km}$ .

<sup>1</sup>P. A. Cloutier, M. B. McElroy, and F. C. Michel, J. Geophys. Res. 74 (26), 6215 (1969).

<sup>2</sup>Sh. Sh. Dolginov, E. G. Eroshenko, and L. N. Zhuzgov, Dokl. Akad. Nauk SSSR 207, 1296 (1972).

<sup>3</sup>M. A. Kolosov and N. A. Savich, in this issue of Usp. Fiz. Nauk, p. 370.

<sup>4</sup>I. S. Hogan, R. W. Stewart, and S. I. Rasool, Radio Sci. 7, 525 (1972).

<sup>5</sup>A. I. Stewart, J. Geophys. Res. 77, 54 (1972).

<sup>6</sup>C. A. Barth, C. W. Hord, A. I. Stewart, and A. L. Lane, Science 175 (4019) 309 (1972).

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