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NASA TM 75717

ON THE STRUCTURE OF THE UPPER ATMOSPHERE OF MARS
ACCORDING TO DATA FROM EXPERIMENTS ON THE
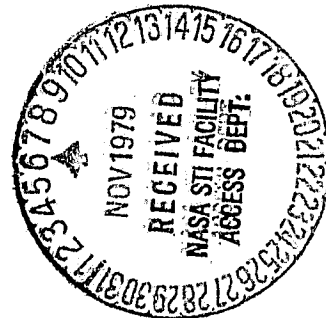"VIKING" SPACE VEHICLES

M. N. Izakov

ON THE STRUCTURE OF THE UPPER ATMOSPHERE OF MARS ACCORDING TO DATA FROM EXPERIMENTS ON THE "VIKING" SPACE VEHICLES

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Altitude profiles of the concentrations of the atmospheric components measured by the on-board mass spectrometers during the descent of the "Viking" lander can be satisfactorily described by assuming that temperature has a smoother profile, and the eddy mixing coefficients are smaller at altitudes of 120 to 170 km than those determined by the authors of the experiments in [1, 2].

The eddy mixing coefficient in the Martian thermosphere during the "Viking" experiments was equal to $5 \cdot 10^7$ cm^2/sec according to our interpretation.

1. During the descent of the lander of the "Viking-1-2" /3*
space vehicles into the Martian atmosphere, excellent measurements were made by means of mass spectrometers. These data were published by Nier and McElroy and their colleagues [1, 2] together with the data from measurements in the lower atmosphere [3, 4]. This information constitutes a major contribution to our knowledge of the structure of the Martian atmosphere. Using these data, investigations of the composition of the atmosphere of Mars and the processes occurring in it have been undertaken [5, 6].

However, in our opinion, certain details in the processing of the data discussed in [2] require further analysis, and certain conclusions are controversial. It is in connection with these matters that the present review was undertaken [5, 6].

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*Numbers in margin indicate pagination of original foreign text.

2. The altitude profiles of the temperature $T(h)$, which were computed in [2] from the altitude profile of the concentration of carbon dioxide gas $n_{CO_2}(h)$, measured by a mass spectrometer, have a meandering form which, in the opinion of the authors, reflects the actual variations of the temperature associated with the passage of tidal waves in the atmosphere. To some extent, this can be the case; however, this can also be due in part to errors in measuring the concentrations and calculating the temperature.

The error in the temperature determined from the concentration profile is $\delta T \approx$ 60° at an altitude of 200 km, according to the authors' estimate [2]. Then it quickly decreases with the altitude, and at altitudes from 140 to 160 km, $\delta T \approx$ ± 5 - 10°, and then it again increases to $\delta T \approx$ 20 - 30° at 120 km altitude. To achieve such a high degree of accuracy in the measurement of temperature at altitudes of 140 - 160 km, it is necessary to measure the concentrations with a precision of $\delta n \approx$ 3 - 5%. Is this possible in flight? In [2] it is assumed that it is possible to determine δn from various concentrations of a single component, determined from various mass peaks (for CO_2 -- 44, 22, 12). Here it is implicitly assumed that, as compared to calibration error, it is possible to neglect: errors due to instability in the characteristics of the apparatus in flight (in particular, the source of ions); errors due to calculating the concentrations in the atmosphere from the concentration in the equipment (the uncontrollable variations in the angle of attack may contribute to this error); and for components falling on some peaks, such as N_2 and CO, errors due to supplementary calculations to isolate them. But if all these errors are actually negligible, then such a strong, non-monotonic dependence of errors on altitude is not understandable. A supplementary analysis of these questions, which only the authors of the experiment can carry out, would be

extraordinarily valuable. It is highly improbable that the error in measuring the concentrations in flight can be markedly less than 25 - 30%.

When calculating the temperature from the concentration, using the barometric formula, was done in [2], the temperature is determined essentially as the logarithmic derivative of the concentration, but during numerical differentiation of an empirical curve containing an error, the error in the result may be increased. From this point of view, it is more advisable to integrate the concentration curve numerically, obtaining the pressure, as was done in [3], and to obtain the temperature from the pressure. When a sufficiently precise computation is performed by this method, the relative error in the temperature may remain the same as the relative error in the concentration.

When differentiating as well as when integrating the profile of $n_\alpha(h)$, it is necessary to interpolate between the experimental points, since they are far apart. Knowing the error in determining the concentration δn_α , it is advisable /5 to construct a smoothed approximation curve (e.g., by the method of least squares), which deviates from the experimental points by not more than the magnitude of the error, rather than to use interpolation. This can help to exclude artificial waves from the curve $n_\alpha(h)$.

The profile of $n_{CO_2}(h)$, computed with constant temperature relative to the height, deviates from the experimental points by not more than 30% (Fig. 1), except for the lower points at $h <$ 130 km, which may have been influenced by the "saturation pressure" in the apparatus [2], and the upper point at 199 km relative to "Viking-2." Therefore, we shall include the fact that at the time of the "Viking-1" experiment, the temperature

of the Martian thermosphere was $T_\omega = 180 \pm 40 \cdot K$, and at the time of the "Viking-2" experiment -- $T_\omega = 130 \pm 30\ K$.

If the structure of the profiles of $n_\alpha(h), T(h)$ actually reflects the influence of the atmospheric waves, then the smoothing curve should be close to the equilibrium profile in the absence of waves -- the most important characteristic of the atmosphere, which must be known first of all.

3. In [2] the very large values, $K = (1 - 4) \cdot 10^9\ cm^2/sec$ at an altitude of 170 km (Fig. 2) are obtained from the profiles of the argon and nitrogen concentrations, $n_{Ar}(h)$, $n_{N_2}(h)$, and the temperature $T(h)$ calculated from the profile of $n_{CO_2}(h)$

In our opinion, the values of K contain a very large error, and at altitudes of 140 - 170 km, this error drastically increases for the following reasons. In [2] K is found by means of the following formula:

$$\frac{\partial \ln n_\alpha}{\partial h} + \frac{\partial \ln T}{\partial h} = -\frac{1}{(D_\alpha + K)}\left(\frac{D_\alpha}{H_\alpha^o} + \frac{K}{\overline{H}}\right), \qquad (1)$$

which is the barometric formula in differential form for the transition layer between the homosphere and the heterosphere. Here

$$\frac{\partial \ln n_\alpha}{\partial h} + \frac{\partial \ln T}{\partial h} = -\frac{1}{H_\alpha}$$

where H_α is the altitude scale for the pressure of the α th component; $H_\alpha^o = \frac{R_o T}{M_\alpha g}$ is the altitude scale for the pressure of the α th component during diffusion equilibrium in the heterosphere; $\overline{H} = \frac{R_o T}{\overline{M} g}$ is the mean altitude scale

during diffusion equilibrium in the homosphere[*]). Clearly, in the homosphere where $K \gg \mathcal{D}_\alpha$, and in the heterosphere where $\mathcal{D}_\alpha \gg K$, formula (1) is converted into the usual barometric formula for H or H_α, respectively. Formula (1) is obtained from the continuity equation for a long-lived atmospheric component (i.e., a component for which the characteristic time of chemical reactions is much greater than the characteristic time of diffusion), if it is assumed that it is in a state of stationary distribution and if we neglect thermal diffusion (cf., e.g., [7]). However, if acoustical-gravitational waves are propagated in the atmosphere, then the distribution deviates from the equilibrium distributions, and it is impossible to neglect the time derivative $\dfrac{\partial n_\alpha}{\partial t}$ in the continuity equation, and in the case of large temperature gradients, it is probably impossible to neglect the thermal diffusion term also, and therefore formula (1) is not applicable. Nevertheless, let us assume that the measured profiles of $n_\alpha(h)$, $T(h)$ are close to the equilibrium distributions. Then (and only then) is formula (1) valid. However, here K as found from (1) has a large error. Indeed, if we solve equation (1) for the desired K, we obtain

$$K = \mathcal{D}_\alpha \frac{\dfrac{1}{H_\alpha^\circ} - \dfrac{1}{H_\alpha}}{\dfrac{1}{H_\alpha} - \dfrac{1}{H}} \qquad (2)$$

[*)] Various authors differ in their use of the term "altitude scale." It is reasonable to define it as the altitude interval over which some quantity undergoes an e-fold change, i.e., $\dfrac{1}{H_\varphi} = \left| \dfrac{1}{\varphi} \dfrac{d\varphi}{dh} \right|$. Here it is obvious that the altitude scales for pressure and concentration are different: $\dfrac{1}{H_s} = \dfrac{1}{H_k} - \dfrac{\partial \ln T}{\partial h}$. However, often, when speaking of an altitude scale, we have in mind an altitude scale for pressuring during diffusion equilibrium H_p°, $H_p^\circ = \dfrac{R_\omega T}{Mg}$.

We see that K is defined by a fraction whose numerator and denominator contain differences, each of which is encumbered with a significant error (since they contain the logarithmic derivative of the profile $n_\alpha(h)$, the temperature calculated from the profile, and the logarithmic derivative of the temperature); moreover, the altitude scales occurring in (2) do not drastically differ in value since the state is approximately the equilibrium state, and the region under consideration is not far from the homopause (where, by definition, K = $\mathcal{D}_\alpha$). It is completely obvious that the resulting error in the determination of K may be very large. But that is not all: to the error due to the structure of the expression for K and the errors contained in its parameters is added the error resulting from the imprecise value of the mean molecular weight $\overline{M} = \frac{\Sigma n_\alpha M_\alpha}{\Sigma n_\alpha}$, which is contained in $\overline{H}$. Indeed, in the range of altitudes under consideration, atomic oxygen becomes one of the basic components of the Martian atmosphere whose concentration is not reliably measured by the mass spectrometer, but is determined (with a certain error) from models. According to the models [6, 8], the concentration n_0 is compared with the concentrations n_{CO_2} at 180 - 200 km altitudes.

We may assume that the errors mentioned mean that in the calculations presented in [2], the values of $\mathcal{D}_\alpha$ "drag after them" the values of K: indeed, it is clear from Fig. 2 that the function K(h) at altitudes greater than 130 km is very similar to the well-known function $\mathcal{D}_\alpha(h)$.

There is one further important circumstance. It is evident from (1) that if the molecular weight of some component M_α is equal to the average molecular weight $\overline{M}$, and therefore $H_\alpha^o = \overline{H}$, then the equation degenerates, and it is impossible to determine K from it. But if M_α is close to $\overline{M}$, then again it

is practically impossible to determine K, since very large differences in K yield very small differences in $n_\alpha(h)$. Such is precisely the case for argon, since M_{Ar} = 40 is close to $\bar{M}$ = 43.5. Therefore, it is advisable to find K from the curve $n_{Ar}(h)$: it is better to use this curve to find $T(h)$; it is true that the error in the homosphere is somewhat larger than with respect to $n_{CO_2}(h)$ due to the greater difference between the molecular weight and the mean molecular weight.

Summarizing, we can say that in order to obtain a reliable determination of the eddy mixing coefficient K by the method under consideration it is necessary, first of all, to use the profile of the concentration of a component with molecular weight which differs markedly from the mean (the profile of helium would be ideal) and, secondly, to have simultaneously reliable measurements of atomic oxygen.

4. To test the ideas discussed above we made such a calculation. In accordance with [1], we took the values of the concentrations of the components n_α at some initial altitude h_o near the lower boundary of the region of the measurements and $n_\alpha(h)$ were computed from the barometric formula, expressed in the following form:

$$n_\alpha(h) = n_\alpha(h_o) \frac{T(h_o)}{T(h)} \exp\left[-\int_{h_o}^{h} \frac{1}{D_\alpha + K}\left(\frac{D_\alpha}{H_\alpha^o} + \frac{K}{H}\right) dh\right]. \qquad (3)$$

Here the coefficients of diffusion $D_{(\alpha, CO_2)}$ were calculated in accordance with the data of [9], where considerable material from laboratory measurements is generalized, and the function $D_{\alpha\beta} = A_{\alpha\beta} T^{-\delta_{\alpha\beta}} e^{-B_{\alpha\beta}/T} n^{-1}$ (where $A_{\alpha\beta}$, $\delta_{\alpha\beta}$, $B_{\alpha\beta}$ -- constants for the given pair of molecules, cf. the table) is proposed.

TABLE

Parameters for calculating the binary coefficients of diffusion of atmospheric gases in CO_2 from the formula:

$$\mathcal{D}_{\alpha\beta} = A_{\alpha\beta}\, T^{b_{\alpha\beta}}\, exp\left(-\frac{B_{\alpha\beta}}{T}\right) n^{-1}$$

$\alpha - \beta$	$A_{\alpha\beta} \cdot 10^{-17}$	$b_{\alpha\beta}$	$B_{\alpha\beta}$
$CO - CO_2$	0,424	0,603	–
$O_2 - CO_2$	1,15	0,661	61,3
$N_2 - CO_2$	2,28	0,570	113,6
$Ar - CO_2$	1,23	0,646	89,1
$H_2 - CO_2$	2,30	0,750	11,7
$He - CO_2$	2,45	0,720	–

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The eddy mixing coefficient K was chosen within limits determined by the theoretical estimates in [10], and then revised in the process of calculation.

Several alternate versions were computed: a) with a smooth profile for $T(h)$ and our value of K; b) with profiles of $T(h)$ and $K(h)$ from [2]; c) with smooth profiles of $T(h)$ and $K(h)$ from [2] (auxiliary version to clarify the influence of K on $n_\alpha(h)$).

Comparison of the profiles (cf. Fig. 1) obtained in versions "a" and "b" show that the profiles of $n_\alpha(h)$, obtained /10 with the smooth profile of $T(h)$, and independent of the altitude of $K = 5 \cdot 10^7$ cm^2/sec, better describe the experimental data. The values of K may decrease sharply at altitudes greater than 150 - 170 km, i.e., a turbopause may occur, just as on Earth. Here, always, $D_\alpha \gg K$; therefore, the values of K have practically no influence on the profile of $n_\alpha(h)$.

A comparison of "b" and "c" shows the extent to which there is little difference between the profiles of $n_\alpha(h)$ when there is a large difference in the values of K, especially in the case of argon, as follows from the considerations presented above.

5. Let us compare the data obtained from the "Vikings" with the data from other experiments and models. According to the interpretation of the measurements on the "Vikings" discussed above, the most probable temperature of the upper Martian thermosphere in the morning (approximately 9 o'clock local time) is $T = 130 \pm 30$ K, and during the day (approximately 16 o'clock local time) is $T = 180 \pm 40$ K. If we sum up the experimental data from which T_∞ can be determined (from the altitude scales for hydrogen, carbon dioxide gas and plasma), then, as we showed

previously [11, 7], there is an obvious explicit dependence of T_∞ on the level of solar activity. When solar activity is high, corresponding to a decim.ter solar radio-radiation flow for a 10.7 cm wave and $F_{10.7} = (150 - 200) \cdot 10^{-22}$ w/m^{2}Hz, the global mean temperature of the upper thermosphere lies within the limits of T_∞ = 350 - 420 K; when solar activity is moderately low, corresponding to $F_{10.7} = 100 \cdot 10^{-22}$ w/m^2 Hz, T_∞ = 270 - 280 K; it is true that this dependency is masked at certain periods, obviously, as the result of the influence of supplementary thermal heat sources in the thermosphere which may be the consequence, on the one hand, of the dynamic influence of the lower atmosphere through the upward distribution and dissipation into the thermosphere of acoustic-gravitational waves, and on the other hand, due to the trans-mission of energy from the solar wind through the magneto-sphere. In every case, the theoretical model [12], in which only the basic source of heat due to the solar ultraviolet, the infrared heat flow and the elimination of heat by molecular conductivity, are taken into account, yields T_∞ close to the empirical results [11]. Consequently, absorption of solar ultraviolet is actually the basic source of heat in the lower latitudes of thermosphere in quiet periods.

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At the time of the experiments on "Viking-1" and "Viking-2," the level of solar activity was extremely low, correspond-ing to $F_{10.7} = 70 \cdot 10^{-22}$ w/m^{2}Hz [13]. If this is taken into consideration, as well as the fact that Mars was close to aphelion (which reduces the solar current by approximately 14% in comparison to periods when Mars was at an average distance from the sun), then, according to the theoretical model [12], we find that in this period, T_∞ should range from 200 to 220 K.

We see that nevertheless * is markedly higher than T_∞

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*)   Illegible in original foreign text.

from the "Viking" data. How can the still colder thermosphere be explained? In our opinion, by the turbulent withdrawal of heat downward from the lower thermosphere. Turbulence in the lower thermosphere provides, on the one hand, a source of heat due to the dissipation of turbulent energy, and on the other, it increases the removal of heat due to turbulent thermal conductivity [7, 14, 15]. To a significant degree, these effects are compensated for; however, under certain conditions, one of them may predominate. This depends on the magnitude of the critical Richardson number which has not yet been sufficiently studied, so that contradictory opinions are expressed about the predominance of one of the effects mentioned [14, 15].

As described above, from the "Viking" results in [2], a rapid growth in K with altitude is obtained: from $(2-4) \cdot 10^7$ cm^2/sec at 100 km altitude to $(1-4) \cdot 10^9$ cm^2 sec at 170 km (Fig. 2). At the same time, according to our interpretation, these same data are satisfactorily described in the case of K at these altitudes -- not changing with altitude and such that $K = 5 \cdot 10^7$ cm^2/sec.

Let us compare these conclusions with other data concerning the eddy mixing coefficient in the upper atmosphere of Mars.

According to our estimate (within the framework of the theory of similarity with the help of the Richardson-Obukhov law), near the turbopause on Mars, K can change within the limits $K = (0.9 - 8) \cdot 10^7$ cm^2/sec [10]. Estimates by a similar, but somewhat different method in [16], gave values close to our lower limit $K = 8 \cdot 10^6$ cm^2/sec. Using our estimates [10] and the estimates of Golitsin [17], according to which at the surface of Mars, $K = (0.1 - 4) \cdot 10^6$ cm^2/sec, and interpolating between them (assuming the linear dependence of $log K$ on h), we

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constructed the profile of $K(h)$ shown in Fig. 2. The homopause, according to our data, lies at altitudes 120 - 140 km. Using these profiles, a model of the composition of the Martian atmosphere was constructed [8], the different versions of which satisfactorily describe the variations in the concentrations of certain components which were measured experimentally. On the other hand, the model of Lin and Donahue [18], in which the value $K = 4 \cdot 10^8$ cm^2/sec was used, does not depend on altitude and gives reduced values of the concentration of O and CO near the ionosphere maximum in comparison with the experimental values. In [19] the profile of $K(h)$ (also represented in Fig. 2) is presented without discussion, giving much larger values of K than ours. Our attempt to calculate a model of the composition [8] makes it possible to conclude that these values of K also are overestimated.

Of course, the profile of $K(\overline{h})$ recommended by us is very approximate. Thus, for example, probably at some altitude below the homopause (at 150 - 170 km), there exists, just as on Earth, a turbopause below which K sharply decreases, since on Mars as on Earth, turbulence in the stable lower thermosphere is maintained due to the energy of the acoustic-gravitational waves arriving from below and dissipating into the lower thermosphere [20-22]. However, this has practically no influence on the profiles of the concentrations of the components, since here under any conditions $\frac{1}{2} \gg K$.

6. Everything discussed above makes it possible to draw the following conclusions.

The convolutions in the altitude profiles of the concentrations of the atmospheric components and the temperature of the atmosphere may reflect, to some extent, the influence of acoustic-gravitational waves and, to some extent, errors in

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measurements and calculations. In order to isolate the real and the artificial waves, a detailed analysis of the methodo- logical errors is necessary. The profiles of the temperature at the time of the "Viking" experiments may have been smoother than the profiles determined in [2].

It is possible to determine the eddy mixing coefficient from the profiles of the concentrations only when a number of conditions are satisfied: stationary distributions of the concentrations; a sufficiently large difference between the molecular weight of the component used and the mean molecular weight (best of all is to use helium); a sufficiently precise determination of the concentrations of all components, includ- ing atomic oxygen. The values of the eddy mixing coefficient K used in [2] are close to the real values at altitudes of 100 - 120 km, but at higher altitudes they are too high, and the error increases with the altitude. At the time of the "Viking" experiments, $K = 5 \cdot 10^7$ cm^2/sec at altitudes of 100 - 150 km, but at higher altitudes the values of K probably de- creased.

The low temperatures of the thermosphere at the time of the "Viking" flights (130 ± 30 K at 9 o'clock local time, and 180 ± 40 K at 16 o'clock) are explained primarily by the low level of solar activity, and to a lesser extent by the fact that Mars was close to aphelion, and also possibly to some degree by the turbulent removal of heat from the thermosphere.

Measurements of the vertical profiles for the concentra- tions of the atmospheric components, presented with sufficient completeness and precision, are some of the most powerful methods for studying the structure and the dynamics of a planetary atmosphere.

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CAPTIONS FOR THE FIGURES

Fig. 1. Height Profiles for the concentrations of components of the Martian atmosphere:

1a) -- from "Viking-1" data;

1b) -- from "Viking-2" data.

Experimental points [1]:

$\odot$ - CO_2, $\triangle$ - Ar, $\square$ - N_2, $\diamondsuit$ - O_2.

Results of calculations under the following conditions:
a) smooth profile of $T(h)$, $K = 5 \cdot 10^7 cm^2/c$;
b) profiles of $T(h)$, $K(h)$ from [2];
c) smooth profiles of $T(h)$, $K(h)$ from [2];

Fig. 2. Altitude profiles of the eddy mixing coefficient $K(h)$ and the coefficient of molecular diffusion $D_{N_2 \cdot CO_2}(h)$ in the atmosphere of Mars.

$\odot$ -- the values of K from the results of "Viking-1" according to [2];

$\triangle$ -- the values of K from the results of "Viking-2" according to [2];

- - - - -- the value of K which satisfactorily describes the "Viking" data according to the present paper;

—·—·— -- the limits of the possible variations in K in the Martian atmosphere according to [8, 10];

—··—·— -- the function $K(h)$, recommended in [19];

—×—×—.— a) ⌐   the function of the coefficients of
             ⌡   molecular diffusion;
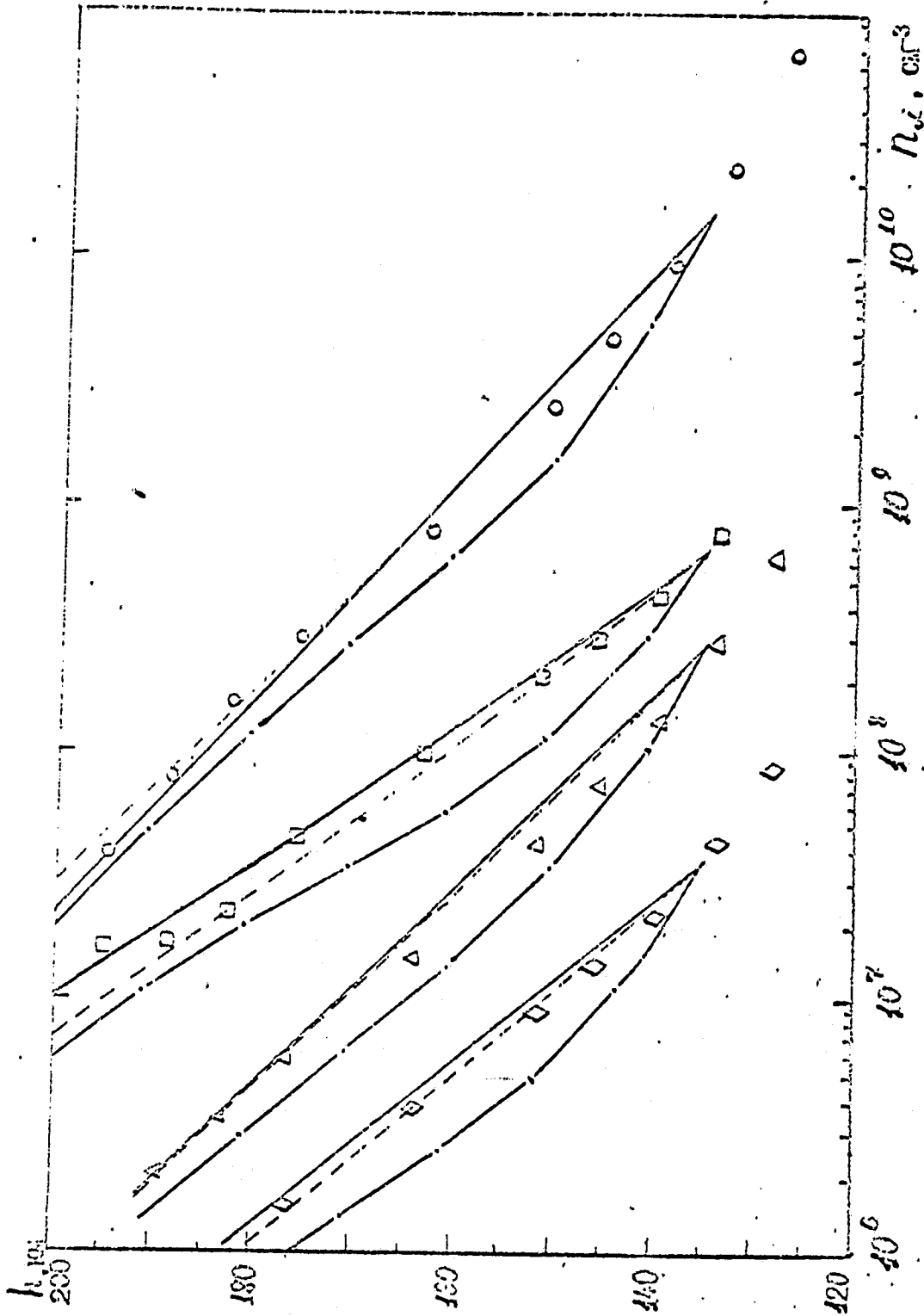———— b) ⌡   (a) - "Viking-1", (b) - "Viking-2."

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Fig. 1a

Fig. 1b

Fig. 2