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A MODEL MARTIAN ATMOSPHERE AND IONOSPHERE

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Summary

The results of the Mariner IV experiment indicate that CO_2 is probably the only significant constituent of the lower Martian atmosphere. Assuming this, we have derived an equilibrium model atmosphere and ionosphere of Mars for the subsolar region and at sunspot maximum. A temperature profile resulted from consideration of radiation balance between the troposphere and mesosphere, vibrational relaxation of CO_2 and the heat released by photodissociation and ionization processes. Dissociation begins at about 70 km producing maximum densities of the dissociation products between 80 and 90 km. Owing to the low density of the atmosphere, CO_2 is distributed in diffusive rather than chemical equilibrium. The photochemistry of the ionosphere includes a multiplicity of charge transfer reactions. The resulting ionosphere is bimodal, the lower peak being predominantly O_2^+. The upper peak, predominantly O^+, is due to the transition between chemical and diffusive equilibrium and may thus be considered an F_2 peak. The difference between this ionosphere and that observed by Mariner IV can be resolved by considering that the F_2 peak is lacking at high latitude due in part to an expected decrease in temperature and in part because O^+ decreases much more rapidly than O_2^+ with decreasing ionizing flux density.

I. Introduction

Thus far no satisfactory model of the Martian atmosphere and ionosphere has been advanced. Early models (i.e. prior to 1964) (Kuiper 1952; Grandjean and Goody 1955; Hess 1958; Kellogg and Sagan 1961; Chamberlain 1962) all included abundant N_2 to account for a considerably overestimated surface pressure. It appears from the results of Mariner IV that retention of N_2 as a significant constituent (e.g. McElroy et al 1965; Chamberlain and McElroy 1966) can be justified mainly by tradition. Several models have recognized the importance of chemical equilibrium in the photodissociation region but have not treated concurrently the diffusion processes. The resulting models are either purely in chemical equilibrium (e.g. Chamberlain and McElroy 1966) or purely in diffussive equilibrium (e.g. Gross et al 1965) neither of which is satisfactory. Many models to not give adequate consideration to heat balance as a determinant of temperature distribution in the atmosphere; in the resulting models the radiation efficiency is not sufficient to dispose of the heat input to the atmosphere (e.g. Johnson 1965). In addition to the obvious atmospheric processes mainly below the mesopause, an adequate treatment of the atmosphere must take into consideration the following: photodissociation; chemical equilibrium; diffusive equilibrium; photoionization; photochemistry of the ionosphere; loss of heat by radiation and conduction; and the resulting temperature profile, which in turn influences each of the preceeding processes. In addition, it is necessary for the equation of continuity to be satisfied for the atmosphere as a whole. In this paper we have tried to consider the simultaneous effect of these phenomena.

II. The Lower Martian Atmosphere

Kaplan, et al (1964), confirmed by Owen (1964), obtained a CO_2 abundance of 50 ± 20 m atm, and their analysis led to a surface composition that can be described empirically by

$$p^{1/2} \, p_1 = 19$$

where p_1 is the partial pressure of CO_2 in mb and p is the total pressure in mb including a few mb of A and N_2. The minimum surface pressure would then be 7 mb, for a pure CO_2 atmosphere.

Observations made from Mariner IV during the flyby of July 15, 1965, (Kliore, et al, 1965) were interpreted to mean a surface scale height of 8.5 km, and a surface pressure of 4 - 6 mb for pure CO_2 or 5 - 7 mb of an equal mixture of CO_2 and A. The small scale height would seem to rule out any significant amount of N_2, or else imply a surface temperature radically lower than for an airless rotating planet of Mar's albedo at its distance from the sun.

Considering probably uncertainties of observations, it appears that the lower Martian atmosphere may be entirely composed of CO_2 with only traces of such other gases as A or N_2.

The mean daytime surface temperature of Mars is about 240°K and the temperature variation from subsolar regions to the terminator has been estimated as about 80°K, while the subsolar temperature has been measured as high as 300°K. A mean subsolar temperature of 280°K and a diurnal range of 160°K are consistent with the observed mean surface temperature. The surface temperature at a solar zenith angle of 70° would then be about 230°K and the air temperature about 50°K less (Mintz 1961) corresponding to the Mariner IV observation.

We shall assume a surface atmospheric temperature of 230°K and a surface pressure of 7 mb, leading to a surface $[CO_2]$ equal to 2.3×10^{17} cm^{-3}[1]. For a pure CO_2 atmosphere, the adiabatic rate is -5.2 deg km^{-1}. Elementary radiation balance consideration put the tropopause temperature at 167°K, and consequently its height at about 12 km, with a pressure of 2 mb. Following the reasoning of Goody (1957) the temperature is assumed to fall off to about 153° well above the tropopause.

Pressure and density calculations on this model put the level of vibrational relaxation for CO_2 at about 60 km, where $[CO_2] = 4 \times 10^{14}$ cm^{-3}. On the assumption that radiative heat transport is negligible above this region, and that no heat source is present, an adiabatic lapse rate is once again assumed, up to the level where $[CO_2] \simeq 10^{14}$ cm^{-3} (about 70 km) and its dissociation begins.

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(1) Throughout this paper, square brackets are used to indicate number densities.

Some of the energy absorbed above this level is converted into heat, all of which can be disposed of by radiation of CO_2 between 70 and 95 km. Nevertheless the dissociation of O_2 and CO_2 above this level constitutes a heat source which affects the temperature profile.

III. Photodissociation Region

The photodissociative solar flux at any level depends upon the optical thickness of the atmosphere through which it has passed. By considering the dissociation cross section, σ, of CO_2 and its diffusion coefficient we concluded that CO_2 is always in diffusion. The optical thickness, τ_{CO_2}, at any height above 95 km is therefore given by
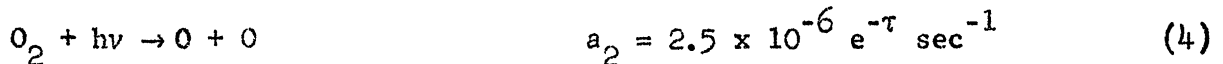
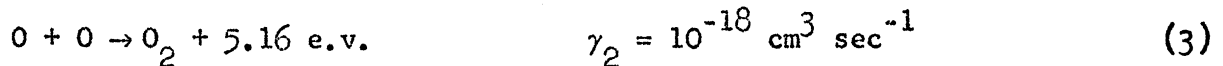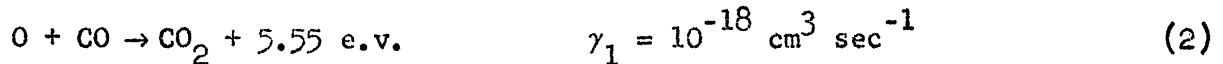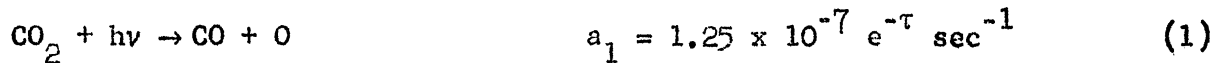$$\tau_{CO_2} = (\sigma \, n \, H)_{CO_2}$$

where

n = number density

and H = local scale height calculated at a mean gravity of 342 cm sec^{-2} at 140 km.

A first approximation to the abundances of the photodissociation products can be calculated by solving the photochemical equilibrium equations using CO_2 in diffusion as the only absorbing constituent. The absorption by the resulting O_2 must now be considered in this iterative procedure.

The dissociative solar flux for CO_2 at the distance of Mars is 6×10^{11} photons $cm^{-2} sec^{-1}$ at sunspot maximum with a dissociation rate of $1.2 \times 10^{-7} sec^{-1}$ (Nawrocki and Papa 1961). The dissociation contiuum begins at 1700 Å with a cross section of about $10^{-20} cm^2$ increasing to $10^{-18} cm^2$ at 1350 Å, reaching a minimum of $4 \times 10^{-20} cm^2$ at 1200 Å and increasing to $10^{-17} cm^2$ below 1150 Å. The dissociative flux for O_2 is about 2×10^{12} photons $cm^{-2} sec^{-1}$, with a dissociation rate of $2.5 \times 10^{-6} sec^{-1}$. The weak Herzberg continuum does not contribute appreciably to the total dissociation, except as noted below, compared to the Schumann-Runge bands. The latter begin at about 1800 Å with a cross section of $10^{-20} cm^2$ increasing to $10^{-17} cm^2$ at 1400 Å and then decreasing to $5 \times 10^{-19} cm^2$ below 1250 Å. The absorption by CO_2 and O_2 in the Herzberg region between 1800 Å and 2400 Å is very small.

This radiation will therefore penetrate to lower levels where it is effective in dissociating molecular oxygen. The atomic oxygen thus liberated is available for the oxidation of carbon monoxide. It is also available for the formation and dissociation of ozone; we ignore this process because any heat flux contributed in the lower atmosphere will not alter the number density distribution greatly in the region of our interest.

The following reactions are considered here. The dissociation rate coefficients a_1 and a_2 are derived by integrating the solar flux over all the wavelengths applicable. The values of γ_1 and γ_2 were estimated from the little information currently available on two-body reactions. Only two-body association processes are considered, since the total atmospheric number density in the dissociation region is too small for the corresponding three-body processes to compete.

$$CO_2 + h\nu \rightarrow CO + 0 \qquad\qquad a_1 = 1.25 \times 10^{-7} e^{-\tau} \, sec^{-1} \qquad (1)$$

$$0 + CO \rightarrow CO_2 + 5.55 \text{ e.v.} \qquad\qquad \gamma_1 = 10^{-18} \, cm^3 \, sec^{-1} \qquad (2)$$

$$0 + 0 \rightarrow O_2 + 5.16 \text{ e.v.} \qquad\qquad \gamma_2 = 10^{-18} \, cm^3 \, sec^{-1} \qquad (3)$$

$$O_2 + h\nu \rightarrow 0 + 0 \qquad\qquad a_2 = 2.5 \times 10^{-6} e^{-\tau} \, sec^{-1} \qquad (4)$$

where τ is the total optical depth for the radiation considered. For this purpose, the following mean cross section were used for each constituent:

$$\sigma(O_2) = 1.62 \times 10^{-18} \, cm^2$$

$$\sigma(CO_2) = 2.16 \times 10^{-19} \, cm^2$$

and $\tau = \sigma(O_2) \displaystyle\int_h^\infty [O_2] \, dh + \sigma(CO_2) \displaystyle\int_h^\infty [CO_2] \, dh$

The rate and continuity equations are then:

$$\frac{d}{dt} [CO_2] = \gamma_1 [0] [CO] - a_1 [CO_2] \qquad (5)$$

$$\frac{d}{dt} [O_2] = \gamma_2 [0]^2 - a_2 [O_2] \qquad (6)$$

$$[CO] = [O] + 2[O_2] \tag{7}$$

$$[CO_2]_o = [CO_2] + [CO] \tag{8}$$

where $[CO_2]_o$ is the initial number density of the primitive CO_2 atmosphere.

Since we are now assuming CO_2 is in a diffusive distribution, equation (8) does not apply, i.e. $[CO_2]$ is known as a function of height.

For $\frac{d}{dt} [CO_2] = 0 = \frac{d}{dt} [O_2]$, these equations can be combined to form the equilibrium equations:

$$\frac{\gamma_1}{a_1} [O]^2 + 2 \frac{\gamma_1 \gamma_2}{a_1 a_2} [O]^3 = [CO_2] \tag{9}$$

$$[O_2] = \frac{\gamma_2}{\alpha_2} [O]^2 \tag{10}$$

$$[CO] = [O] + 2 [O_2] \tag{7}$$

For any given value of $[CO_2]$, one may merely solve the cubic equation (9) for $[O]$ and then calculate $[O_2]$ and $[CO]$.

The resulting distiribution is itself a first approximation. Before a final distribution can be ascertained, we must obtain a temperature profile for the upper atmosphere, since all the constituents will be in diffusive equilibrium above a certain level.

IV. The Transition between Chemical and Diffusive Equilibrium

When the diffusive transport rate of an atmospheric constituent is greater than its chemical reaction rate, the distribution of the constituent is governed by diffusive, rather than chemical equilibrium. The transition between the two processes takes place gradually near the lowest level at which a molecule can diffuse upward for one scale height without suffering a sufficient number of collisions to result in a reaction. This condition occurs when the diffusion time constant t_d is comparable to the chemical time constant t_c:

$$t_d = \frac{H^2}{D} \tag{11}$$

$$t_c = (\beta \, n)^{-1} \tag{12}$$

where H is the scale height, D the diffusion coefficient and $(\beta \, n)$ the rate coefficient for the reaction of the given species with another whose number density is n. If we let:

$$D = \frac{C}{n_T} T^{1/2}$$

where C is a constant and n_T is the total number density, and use for H the conventional expression:

$$H = \frac{kT}{mg}$$

the criterion becomes:

$$n \, n_T = \frac{C}{\beta} \left(\frac{mg}{k}\right)^2 T^{-3/2} \tag{13}$$

The classical diffusion coefficient for a molecule of species 1 through a gas of species 2 is

$$D_{12} = \frac{kT}{m_1 \nu_{12}} = \sqrt{\frac{\pi k T}{8 m_1} \left[\frac{m_2}{m_1 + m_2}\right]} \cdot \frac{1}{n_2 \sigma_{12}} \tag{14}$$

where m_1 and m_2 are the molecular masses, ν_{12} is the collision frequency of species 1 with species 2 and σ_{12} is the gas-kinetic cross section for collision between the two species. The binary diffusion coefficient differs from that derived by Chapman and Cowling (1960) in that the square bracket in the radical is inverted and a slight difference in the constants. Values in the International Critical tables give a $T^{3/4}$ dependence rather than the classical $T^{1/2}$. Evaluation of the various expressions for diffusion coefficients for the temperatures of interest give results that are roughly equal. For diffusion through a mixture of different gases, including self-diffusion, we take the coefficient for species 1 as:

$$D_1 = \sqrt{\frac{\pi k T}{8 m_1}} \left[\sum_s n_s \sigma_{12} \sqrt{1 + \frac{m_1}{m_s}} \right]^{-1} \tag{15}$$

If μ is the molecular weight of the diffusing species, and g is assumed as 342 cm sec^{-2} for a height of about 140 km in the Martian atmosphere,

$$H^2 = 5.914 \times 10^{10} \left(\frac{T}{\mu}\right)^2 \text{ cm}^2$$

The chemical time constants t_c for the pertinent reactions are as follows, using the rate coefficients of reactions (1) to (4).

For CO the only loss mechanism is reaction (2). Thus

$$t_c(CO) = \frac{10^{18}}{[O]} \text{ sec}$$

Reactions (2) and (3) provide loss mechanisms for O. Then

$$t_c(O) = \frac{10^{18}}{[O] + [CO]} \text{ sec}$$

The photo-dissociation time constants for CO_2 and O_2 are $8 \times 10^6 \, e^{-\tau}$ sec and $4 \times 10^5 \, e^{-\tau}$ sec, respectively. With the assumed parameters, CO_2 will nowhere be in chemical equilibrium. Above 110 km, all the constituents will be in diffusive equilibrium.

One more condition must be imposed: the equation of continuity for the atmosphere as a whole must be satisfied by the O, O_2 and CO distributions.

$$\int_{70}^{\infty} O \, dh + 2 \int_{70}^{\infty} O_2 \, dh = \int_{70}^{\infty} CO \, dh$$

This neglects any escape of these constituents from the atmosphere. At the bottom of the diffusion level then $[CO] = \frac{4}{7}([O] + [O_2])$

V. Photo-ion Production

The rate of ion production at any level is found by multiplying the number density of each constituent by its respective ionization rate q. The rate of electron production is of course the sum of the ion production rates. Estimation of the q's involves consideration of the solar spectrum (Norton et al 1962) and the absorption cross sections (σ). To facilitate computation, certain wavelengths were grouped together, and the average cross section and solar flux taken. Table I shows the values used. For

each 10 km level for the diffusive region above 110 km the optical depth $\tau(\lambda)$ and $e^{-\tau(\lambda)}$ were computed by summing σ N H for each constituent, where N is the number density and H the local scale height. Below 110 km the τ's were calculated from the density profile. Ionization cross sections were taken from Nicolet et al (1960) and Norton et al (1962). Those for CO_2 were assumed equal to those for CO for want of better information. Finally the q's were obtained by summing $J(\lambda)\ \sigma(\lambda)\ e^{-\tau(\lambda)}$ over all the wavelengths for each constituent, where $J(\lambda)$ is the photoionizing flux at the top of the atmosphere in the given wavelength bands. Figure 3 shows the photoion and electron production as a function of height. Two peaks occur in ion production: a sharp one around 85 km due to the ionization of O_2 and a broad one around 140 km. The latter is due to the variation of cross section and ionizing flux with wavelength and to the fact that the ionizing flux at a given height depends on the total absorption above that height. The sum of the peaks of each constituent produces a broad peak of electron production. The prominant peak due to O_2 ionization is largely attributable to ionizing flux in the wavelength range from 1030 to 912 Å which is not absorbed by the other constituents. A small O peak at 105 km is due to radiation in the region 912 - 885 Å which is not absorbed by either CO or CO_2. It is interesting to note that the large O_2 peak occurs despite the relatively small amount of O_2 in the atmosphere.

VI. Heat Balance

The heat produced by the ionization process is an important input to the atmosphere and together with the heat produced by photodissociation is a primary determinant of the temperature profile of the upper atmosphere. Since we have already assumed a temperature profile to derive the various distributions, we must now iterate the whole procedure until a self-consistent temperature profile is achieved.

The equation of heat transport is:

$$AT^{1/2}\ \frac{dT}{dh} = \int_{h}^{\infty} (E-R)\ dh \tag{16}$$

where $AT^{1/2}$ is the thermal conduction coefficient, E is the rate at which heat is deposited by photochemistry at each level and R is the rate of

radiation at each level. The radiation term R includes radiation by each of the important constituents as given in Bates 1951.

$$R(0) = 1.7 \times 10^{-18} \ [0] \ \exp \ (- \tfrac{223}{T}) \ \text{erg cm}^{-3} \ \text{sec}^{-1} \tag{17}$$

$$R(CO_2) = 3.4 \times 10^{-28} \ [CO_2] \ n \ \exp \ (- \tfrac{990}{T}) \ \text{erg cm}^{-3} \ \text{sec}^{-1} \tag{18}$$

where n = total number density, and

$$R(CO) = 2.8 \times 10^{-23} \ T^2 \ [CO] \tag{19}$$

$R(CO)$ represents rotational rather than vibrational radiation since the latter is a much less efficient process.

The coefficient A is given as 180 erg cm^{-1} sec^{-1} deg$^{-3/2}$ for O_2 and 360 for 0 (Nicolet 1961); for CO it may be estimated as 180 erg cm^{-1} sec^{-1} deg$^{-3/2}$ and for CO_2 as 130 erg cm^{-1} sec^{-1} deg$^{-3/2}$. As a mean value of A at any level we adopted

$$A = \frac{130[CO_2] + 180([O_2] + [CO]) + 360[0]}{n} \tag{20}$$

Using the above equations (17) - (20) and the assumed value of T, we may calculate $\frac{dT}{dh}$, which by integration gives a new T as a function of height. The nature of the dependence of each R on T makes an analytic solution impractical, thus necessitating an iterative procedure. An exhaustive discussion of the radiation processes has been given by McElroy et al 1965.

One of the uncertain quantities is the total heat input to the upper atmosphere resulting from the photochemical processes. The value of Harris and Priester 1962, was reduced for the distance of Mars. Our estimated total of 1 erg cm^{-2} sec^{-1} was divided between photoionization and photo-dissociation by attributing 1.9×10^{-11} ergs/photoion pair produced and 4.75×10^{-13} ergs/photodissociation.

VII. The Ionosphere

With this self-consistent temperature profile and the resulting consistent constituents we are now ready to calculate the ionosphere by combining the reactions of the neutral and ionized constituents. We begin by assuming photochemical equilibrium.

Although only two chemical elements are present, the reactions are about as complex as those in the lower terrestrial ionosphere. Table II gives the reactions considered important and their estimated rate coefficients. Recombination coefficients are called α; in each case except that for $O^+ + e$ they represent dissociative recombination. Only two-body radiative recombination is important for O^+, since the density of the atmosphere is too low for a three-body process.

Some discussion of the assumed rate coefficients is in order. In most cases they are not known to better than an order of magnitude, if that well. The radiative recombination rate α_1 is taken from Nawrocki and Papa (1961) and the dissociative rate α_2 from there and Norton et al (1962). α_3 and α_4 were estimated from Nawrocki and Papa (1961) and analogy to the NO_2^+ and NO^+ rates. The atom-ion reaction is assumed to be mostly an ion-atom exchange rather than a true charge exchange. The rate k_4 is taken from Norton et al (1962); the rate k_9 from Fjeldbo et al (1966). The other k's are estimated largely by analogy, considering the energy balance and the possibility of absorption of excess energy by excitation.

No three-body reactions are included because of the low total particle density assumed; the probable rate of 10^{-27} cm^6 sec^{-1} (Bortner 1965) makes such reactions orders of magnitude less than two-body reactions in the regions we are concerned with.

From the reactions in Table II, the following equilibrium equations can be written using n_e for the electron density:

$$(\alpha_1 n_e + k_4[O_2] + k_9[CO_2]) [O^+] = (q_1 + k_1[CO^+] + k_2[CO_2^+]) [O] + k_3[O_2][CO^+] \quad (21)$$

$$\alpha_2 n_e[O_2^+] = (q_2 + k_4[O^+] + k_5[CO^+] + k_6[CO_2^+]) [O_2] + k_7[O][CO_2^+]$$
$$+ k_9[CO_2][O^+] \quad (22)$$

$$(\alpha_3 n_e + k_1[O] + (k_3 + k_5) [O_2]) [CO^+] = (q_3 + k_8[CO_2^+]) [CO] \quad (23)$$

$$(\alpha_4 n_e + (k_2 + k_7) \, [O] + k_6 [O_2] + k_8 [CO]) \, [CO_2^+] = q_4 [CO_2] \tag{24}$$

$$n_e = [O^+] + [O_2^+] + [CO^+] + [CO_2^+] \tag{25}$$

Dissociative recombination turns out to be an unimportant mechanism for removal of CO^+ and CO_2^+. This makes solution of equations (24) and (23) very simple. Radiative recombination with O^+ is also insignificant because electrons and ions are in diffusion far below the level where the process would manifest itself. The only direct recombination of importance is that of O_2^+.

Since $\alpha_1 n_e$, $\alpha_3 n_e$ and $\alpha_4 n_e$ can be neglected, $[CO_2^+]$ can be calculated directly from equation (24) given the neutral abundances. Next $[CO^+]$ can be calculated from equation (23), since we know $[CO_2^+]$. Now we can calculate $[O^+]$ from equation (21).

We now can write $[O_2^+] = n_e - [O^+] - [CO^+] - [CO_2^+]$ and put this in equation (22) which thereby becomes a quadratic equation in n_e. At this point we can solve for $[O_2^+]$ from equation (25).

The equilibrium equations were solved to obtain values of the number densities of electrons and of the various ionic species as a function of height, assuming chemical equilibrium. It turns out that CO^+ and CO_2^+ are minor ions, the predominant ones being O^+ and O_2^+. O_2^+ is the predominate ion below about 180 km, despite the fact that neutral O is more abundant than neutral O_2. This predominance is due to the multiplicity of charge exchange reactions terminating with the ion O_2^+ which has the lowest energy.

Thus far we have considered the ionosphere to be in chemical equilibrium. A number of references (Bauer 1965; Hanson 1962, etc) have pointed out that chemical equilibrium can obtain only if the diffusion rates are significantly smaller than the chemical reaction rates. There is a critical level above which diffusive equilibrium predominates. Above 180 km the important chemical reactions are those which remove the atomic oxygen ions. The rate of this loss must be compared with the diffusion rate for O^+ in an atmosphere of predominantly O and CO.

The ambipolar diffusion coefficient c^+ o^+ in O or CO is given approximately by

$$D_a = 7 \times 10^{15} \frac{T}{n} \text{ cm}^2 \text{ sec}^{-1}$$

for the temperature we have assumed, where n is the total atmospheric number density (Nawrocki and Papa 1961). The diffusion time constant

$$t_d = H^2/D_a \quad \text{sec}$$

where H is now the effective ion scale height. The chemical reaction rate is given by the coefficient of $[O^+]$ in equation (16). We find that $t_c \approx t_d \approx$ 7×10^3 sec at 340 km. Using this information, the electron-density profile is faired in between the chemical lapse rate and diffusive rate. The resulting ionosphere is shown in Figure 4.

VIII. Discussion

The ionosphere constructed by using the above procedures has three peaks of comparable magnitude; in the lower two O_2^+ is predominant while the upper peak is a "Bradbury" peak of O^+. Study of the equilibrium ion density equations reveals that with decreasing solar flux O^+ decreases much more rapidly than O_2^+. This is largely due to the assumed high efficiency of charge rearrangement between O^+ and CO_2. Although no study has yet been made of a dynamic model, some broad conclusions can be drawn.

In times of low solar activity and for large solar zenith angles, the top peak would disappear and the ionospheric temperature would decrease substantially, leaving a single thin but still rather dense O_2^+ peak. The density of the single peak will depend on the latitude one is considering. At low latitudes, the solar flux decreases steeply with time and a dynamic analysis is necessary. For high latitudes however, the solar flux is never very high, so that a much lower density would be expected. This is indeed consistent with the Mariner IV data which measured the ionosphere at high latitude as well as large solar zenith angle. If the temperature is as low as 170°K, the O_2^+ region would have a scale height of 25 km. It is obvious that a temperature of 85 - 90°K which is hard to reconcile with heat balance considerations is not necessary.

TABLE I

$\Delta\lambda$ $\overset{\circ}{A}$	J $(x\ 10^8)$	O_2		O		CO	
		$J\sigma$ $(x\ 10^{-9})$	σ $(x\ 10^{-17})$	$J\sigma$ $(x\ 10^{-9})$	σ $(x\ 10^{-17})$	$J\sigma$ $(x\ 10^{-9})$	σ $(x\ 10^{-17})$
1030-912	115.8	59.56	.515				
912-885	54.0	39.3	.728	14.5	.269		
885-770	38.4	39.8	1.017	11.7	.305	49.5	1.290
705-200	127.9	217.0	1.700	112.2	.880	157.8	1.234
200-130	23.0	15.2	.661	7.6	.330	13.13	.571
130- 90	6.0	1.0	.1667	.52	.0868	.96	.160
90- 44	14.1	1.18	.0838	.588	.0417	.923	.0655

TABLE II

Reaction	Rate Coefficient
$O + h\nu \rightarrow O^+ + e$	q_1 sec^{-1}
$CO^+ + O \rightarrow O^+ + CO + 0.38$ e.v.	$k_1 = 2 \times 10^{-11}$ cm^3 sec^{-1}
$CO_2^+ + O \rightarrow O^+ + CO_2 + 0.16$ e.v.	$k_2 = 1 \times 10^{-11}$ cm^3 sec^{-1}
$CO^+ + O_2 \rightarrow O^+ + CO_2 + 0.73$ e.v.	$k_3 = 2 \times 10^{-11}$ cm^3 sec^{-1}
$O^+ + e \rightarrow O + h\nu$	$\alpha_1 = 2 \times 10^{-12}$ cm^3 sec^{-1}
$O^+ + O_2 \rightarrow O_2^+ + O + 1.58$ e.v.	$k_4 = 5 \times 10^{-11}$ cm^3 sec^{-1}
$O_2 + h\nu \rightarrow O_2^+ + e$	q_2 sec^{-1}
$CO^+ + O_2 \rightarrow O_2^+ + CO + 1.93$ e.v.	$k_5 = 5 \times 10^{-11}$ cm^3 sec^{-1}
$CO_2^+ + O_2 \rightarrow O_2^+ + CO_2 + 1.71$ e.v.	$k_6 = 5 \times 10^{-11}$ cm^3 sec^{-1}
$CO_2^+ + O \rightarrow O_2^+ + CO + 1.36$ e.v.	$k_7 = 5 \times 10^{-11}$ cm^3 sec^{-1}
$O_2^+ + e \rightarrow O + O + 6.92$ e.v.	$\alpha_2 = 2 \times 10^{-7}$ cm^3 sec^{-1}
$CO + h\nu \rightarrow CO^+ + e$	q_3 sec^{-1}
$CO_2^+ + CO \rightarrow CO^+ + CO_2 + 0.56$ e.v.	$k_8 = 2 \times 10^{-11}$ cm^3 sec^{-1}
$CO^+ + e \rightarrow C + O + 2.85$ e.v.	$\alpha_3 = 1 \times 10^{-7}$ cm^3 sec^{-1}
$CO_2 + h\nu \rightarrow CO_2^+ + e$	q_4 sec^{-1}
$O^+ + CO_2 \rightarrow O_2^+ + CO + 1.20$ e.v.	$k_9 = 1 \times 10^{-9}$ cm^3 sec^{-1}
$CO_2^+ + e \rightarrow CO + O + 8.23$ e.v.	$\alpha_4 = 3 \times 10^{-7}$ cm^3 sec^{-1}
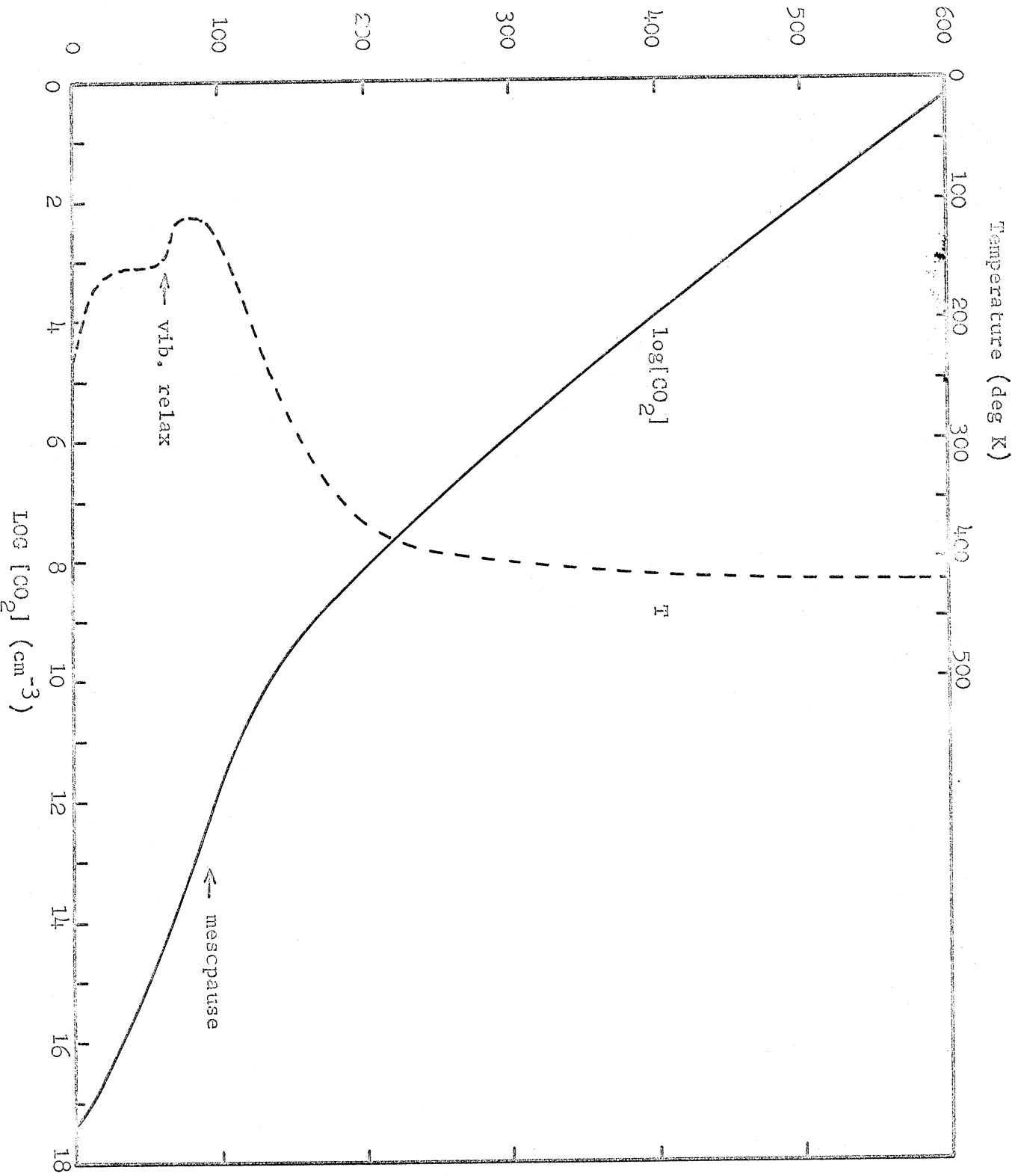
FIGURE 1

FIGURE 2

FIGURE 3

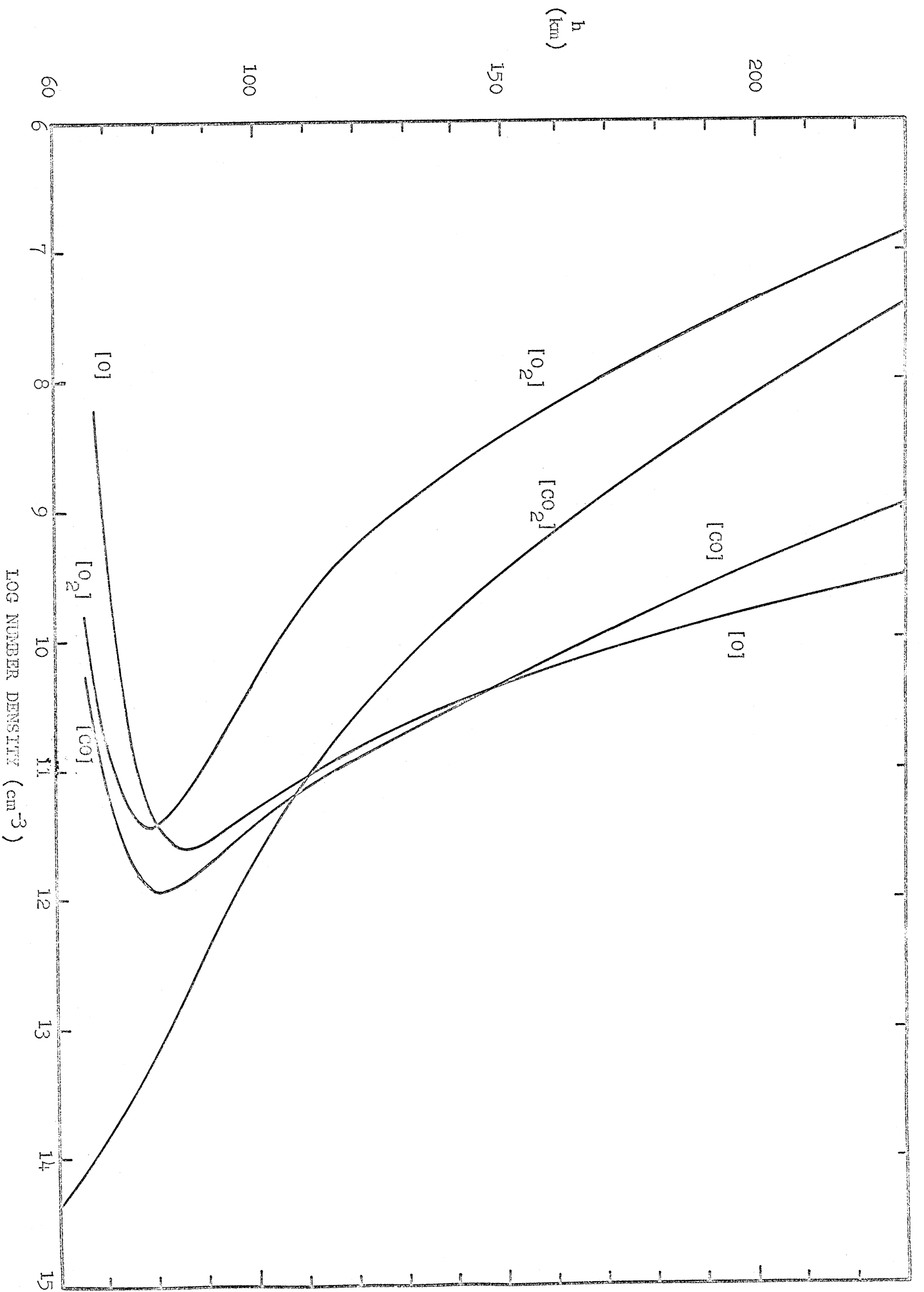ION AND ELECTRON NUMBER DENSITY (cm^{-3})

CO_2^+

$[CO^+]$

$[O^+]$

$[O_2^+]$

n_e
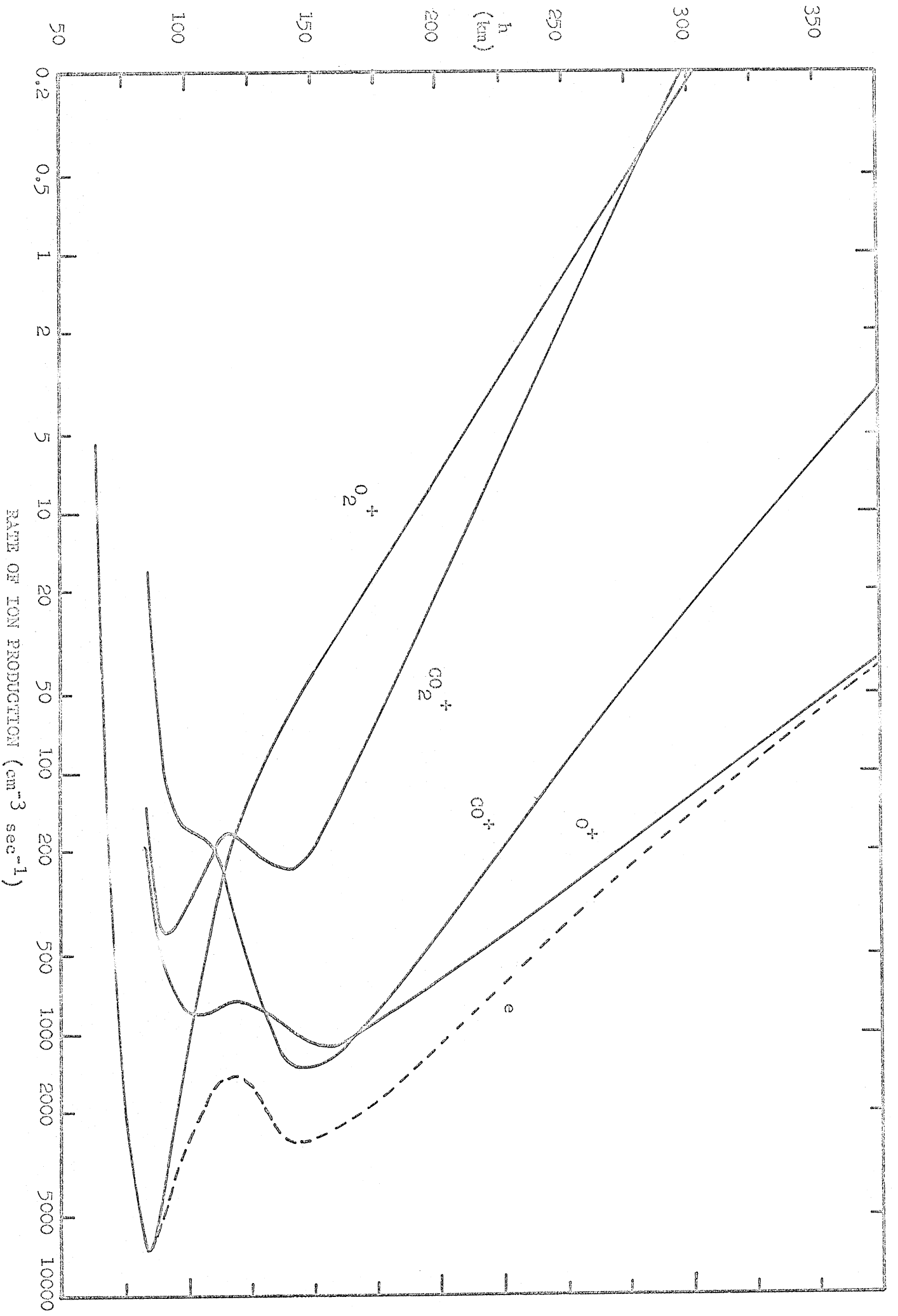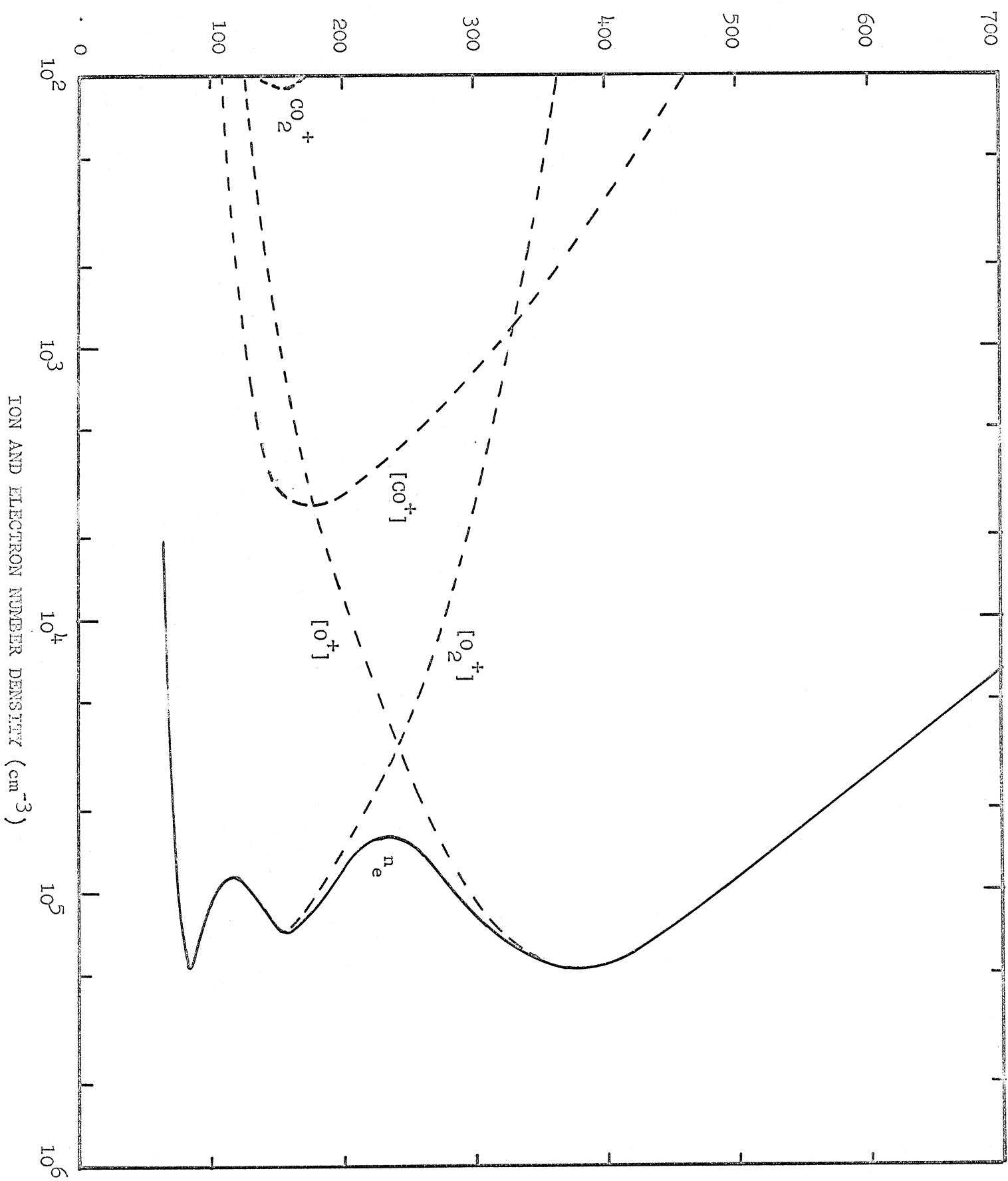
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