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The composition and structure of the terrestrial atmosphere

by M. NICOLET

FOREWORD

The text of this Aeronomica Acta is the lecture given at the Symposium of the Air Force and General Electric on : "Dynamics of Manned Lifting Planetary Entry" which has been held in Philadelphia (USA) from 29th till 31th October 1962. This paper will be published in the book with the other papers presented at his Symposium.

M. NICOLET.

AVANT-PROPOS

Le texte de cet Aeronomica Acta est l'exposé que nous avons fait au Symposium de l'Air Force et General Electric sur : "Dynamics of Manned Lifting Planetary Entry" qui a eu lieu à Philadelphie (USA) du 29 au 31 octobre 1962. Le texte sera publié dans le livre couvrant tous les exposés faits à ce Symposium.

M. NICOLET

VOORWOORD

De tekst van deze Aeronomica Acta is de uiteenzetting welke wij gehouden hebben op het Symposium van de Air Force en General Electric, handelend over "Dynamics of Manned Lifting Planetary Entry". Het werd gehouden van 29 tot 31 oktober 1962 te Philadelphia (USA). De tekst zal verschijnen in een boek welke al de uiteenzettingen omvat welke tijdens dit Symposium gedaan werden.

M. NICOLET.

VORWORT

Der Text dieses Aeronomica Acta ist das Exposé, das wir in Philadelphia (USA, 29. - 31. October 1962) zum Symposium des Air Force und General Electric über "Dynamics of Manned Lifting Planetary Entry" vorgestellt haben. Der Text wird mit den anderen Aufträgen in ein Buch herausgegeben werden.

M. NICOLET.

Table of Contents

ABSTRACT

After a general description of the various regions of the terrestrial atmosphere, an analysis of physical conditions leading to the knowledge of the heterosphere is given. Molecular dissociation and diffusion in the lower thermosphere are introduced to determine the concentration of various constituents (N_2, O_2, O, N, He, H). Ultraviolet heating and conduction are considered as essential processes of production and loss of heat in the thermosphere. Finally, the ionospheric diffusion equilibrium is studied after considering the system of ionic reactions occuring in the ionosphere.

1. Introduction

During the past several years many observations concerning the composition and structure of the upper atmosphere have been made. Solar radiations in the far ultraviolet and X-ray regions have been observed. The distribution of molecular oxygen as a function of altitude has been determined. The density of the atmosphere has been measured by means of sounding rockets to 200 km and by satellites to 1500 km. The state of the ionosphere, studied by means of sondes to the highest altitudes, yields the vertical distribution of electrons. Finally, mass spectrometry observations reveal the existence of ions whose vertical distribution is different from the distributions of the neutral constituents.

There results from this accumulation of data an extensive documentation from which it is not always possible to draw final conclusions. This condition is illustrated by the fact that, often, different interpretations are given for the same observations. Thus, the general picture of the composition and structure of the upper atmosphere differs according to the various investigators. It is convenient to define a suitable set of simple physical parameters and to base the model on these elementary physical conditions in order to obtain a simple picture of the structure of the atmosphere.

2. General Description

Before setting forth the exact conditions of the upper

atmosphere, it is necessary to refer to those which deter-
mine the characteristics of the lower atmosphere (altitudes
up to 50 km). The lower atmosphere is divided into two
regions: the troposphere and the stratosphere. The troposphere
is essentially characterized by the decrease of temperature
with altitude above the earth; which is the source of heat.
The stratosphere, which starts from the tropopause (minimum
temperature), by a layer where the temperature does not de-
crease with altitude but increases in a manner which leads
to a maximum of temperature of the order of $0^\circ C$ at the
stratopause. If the altitude of the tropopause varies with
the latitude (12 $\pm$ 6 km) and depends, in the middle latitudes,
on atmospheric conditions, the same applies to the altitude
of the stratopause (50 $\pm$ 5 km). Moreover, it is clear that,
if the temperature reaches a maximum in the stratopause, a
heat source must be associated with it. This heat results
from the absorption by the atmospheric ozone of ultraviolet
solar radiation.

Above the stratopause, temperature decreases with
altitude to a minimum which may be as low as $160^\circ K$. This
layer, called mesosphere, extends to about 85 $\pm$ 5 km. This
level, corresponding to the mesopause, constitutes the lower
limit of the upper atmosphere. In fact, up to the mesopause,
the composition of the atmosphere is characterized by con-
stant proportions of nitrogen (N_2 = 78 %), oxygen (O_2 = 21%),
and argon (A = 1%). Mixing is the essential characteristic,

and the molecular mass remains constant; thus the name
homosphere.

In the homosphere, the principal constitutents are
subject to a vertical distribution which depends upon the
law of the static atmosphere

$$\frac{dp}{dz} = - g\rho \qquad (1)$$

where p is the pressure, which varies with altitude z,
$\rho = nm$ is the density, corresponding to a concentration n
and a constant molecular mass m, and g is the acceleration
of gravity. Moreover, the pressure is that of a perfect
gas,

$$p = n\,k\,T \qquad (2)$$

where n is the total concentration, T the absolute temperature,
and k the Boltzmann constant.

From equations (1) and (2) the general vertical distri-
bution law of the atmosphere may be written as

$$\frac{dp}{p} = \frac{d\rho}{\rho} + \frac{dT}{T} = - \frac{dz}{kT/mg} \cdot \qquad (3)$$

Thus, one obtains a relationship between pressure, density, and
temperature characterized by the parameter H which is the
scale height,

$$H = kT/mg \qquad (4)$$

with $\qquad \frac{dH}{H} = \frac{dT}{T} - \frac{dg}{g} \cdot \qquad (4a)$

The minor constituents, such as H_2O, O_3, CH_4, N_2O, NO_2, etc., do not necessarily follow the hydrostatic distribution and are subject, in particular in the mesosphere, to photochemical reactions which cause changes in the composition.

Above the mesopause, the physical conditions change. In the first place, molecular oxygen undergoes photodissociation caused by ultraviolet radiation with a wavelength shorter than $\lambda = 2400\overset{o}{A}$, and, in particular, in the region of $\lambda < 1750 \overset{o}{A}$. Secondly, the effect of diffusion becomes more important than the mixing effect and, thus, the vertical distribution of the principal constituents must be modified. Therefore, we reach the heterosphere, where the average molecular mass is no longer constant and (4a) must be replaced by:

$$\frac{dH}{H} = \frac{dT}{T} - \frac{dm}{m} - \frac{dg}{g} \, . \tag{4b}$$

Thus, in defining the gradient of the scale height by

$$\beta = \frac{dH}{dz} , \tag{5}$$

the general equation of the density of the atmosphere, (3) and (5),

$$\frac{d\rho g}{\rho g} = \frac{1 + \beta}{\beta} \frac{dH}{H} \tag{6}$$

depends on the ratio dH/H defined by (4a) in the case of the homosphere and by (4b) in the case of the heterosphere.

The variation of ρg with altitude in the heterosphere indicates, first of all, that between 100 km and 200 km the

increase of H with altitude is such that a rapid increase of temperature is necessary. Here we find the _thermosphere_ which is characterized by large temperature gradients. Specifically, at an altitude of 150 km, according to these conditions, one must find the following extremes:

$T(^{\circ}K)$	$1000^{\circ}K$	$500^{\circ}K$
dT/dz	$20^{\circ}K/km$	$5^{\circ}K/km$

Such gradients can only be explained by a heating due to the absorption of ultraviolet rays with $\lambda < 1000 \overset{o}{A}$. The intensity of this radiation depends upon solar activity and, therefore, with variable heating, the temperature and its gradient must be different according to the conditions of solar activity. They will be highest in the course of maximum solar activity and lowest during minimum solar activity. Moreover, the propagation of heat in the upper atmosphere is essentially effected by conduction, and a tendency towards an isothermal condition must be seen at a certain altitude, defined by the word _thermopause_, above which the atmosphere is practically isothermal.

Finally, the heating during the day by solar ultraviolet and cooling by conduction during day and night leads to a diurnal variation of atmospheric temperature. Thus, there exists a diurnal variation of the density associated with a temperature variation.

Above the thermopause, equation (6) can again be applied, but in place of (4a and b) one writes:

$$\frac{dH}{H} = - \frac{dm}{m} - \frac{dg}{g} \qquad\qquad (4c)$$

indicating that the gradient β depends, essentially, on the variation of the molecular mass m. Thus, above 500 km, that is above the thermopause, diffusion is practically the only effect. The action of diffusion is such that from 100-500 km a minor constituent, such as helium, has its relative abundance increase with respect to the principal constituents (O_2, N_2 and O). This relative increase produces a helium belt in the vicinity of 1000 km. Finally, atomic hydrogen, the lightest element, is dominant at the highest altitudes.

The application of equation (6) cannot, however, be made above a certain limit. In the first place, the effect of centrifugal force must be considered and equation (6) must be corrected to include the effect of the rotation of the atmosphere. Secondly, for the light elements, such as helium and hydrogen, one must take account of their escape from the earth's atmosphere starting in the region where collisions no longer occur. This region, where the effect cf collisions is negligible, is called the exosphere. The absence of collisions, practically, does not change, with regard to the hydrostatic distribution, the vertical distribution of the heavy elements, such as nitrogen and oxygen. One finds in fact for T = constant, m = constant, (4b),

$$\frac{dH}{H} = - \frac{dg}{g} = - \frac{2dr}{r} \qquad\qquad (4d)$$

if r be the distance from the center of the earth. Therefore, for a constituent of mass m_1

$$\frac{\rho_1}{\rho_{1,a}} = e^{-\left(\frac{a}{r}\right)\frac{r-a}{H_{1a}}} \qquad (7)$$

where a is the starting point $r = a$. Equation (7) for the density leads to

$$\int_a^\infty \rho_1 \, dr \simeq \rho_{1a} H_{1a} \qquad (8)$$

with a sufficient approximation if H_r is not too large. This is the case for nitrogen and oxygen but not for helium and hydrogen.

At a very great distance from the center of the earth, in a rotating atmosphere, one must modify (7) by writing:

$$\frac{\rho_1}{\rho_{1a}} = \exp\left[-\left(\frac{a}{r}\right) \frac{r-a}{H_{1a}} \left\{ 1 - \frac{\omega_e^2}{2ag_a} (r + a)r \sin^2 \theta_e \right\} \right] \qquad (9)$$

where ω_e is the angular velocity of the earth and θ_e represents the colatitude.

However, when (8) is not applicable because of the large value of H_{1a}, that is for a value of the ratio r/H approaching unity, one obtains a different formula for the vertical distribution (cf. references [1] to [6]). One can write in place of (7), for example[6],

$$\frac{\rho_1}{\rho_{1a}} = \left\{ \exp\left[-\left(\frac{a}{r}\right) \frac{r-a}{H_{1a}} \right] \right\} \left\{ 1 - \left(1 - \frac{a^2}{r^2} \right)^{1/2} e^{-(a/r)^2 \frac{1}{1+a/r} \frac{a}{H_{1a}}} \right\} \qquad (10)$$

indicating a more rapid decrease of the density than that resulting from the hydrostatic distribution. This law is applicable to helium and even more so to hydrogen at altitudes above 3000 km.

As a consequence of diffusion, the terrestrial atmosphere continues to exist up to the highest altitude in the form of neutral particles (He, H). However, it is necessary to introduce the word ionosphere in order to include the effects of physical conditions adapted to ions and electrons. These are in effect charged particles which come from the neutral constitutents of the atmosphere, subjected to electrostatic conditions and the terrestrial magnetic field. The ionosphere starts at about 50-70 km due to the effect of primary cosmic rays (Nicolet and Aikin, 1960), and the electron concentration is on the order of 100 electrons cm^{-3}. Under the influence of the most penetrating solar radiation, X-rays of $\lambda < 10$ Å and Lyman α to 1216 Å, the concentration reaches a maximum of about 1000 electrons-cm^{-3} in the D region near 85 km.

In the E region there appears a maximum concentration on the order of 100,000 electrons cm^{-3} between 100 and 120 km as a result of the absorption of X-rays from 10 to 100 Å and of ultraviolet rays with $\lambda > 910$ Å. Finally, the principal peak, corresponding to a concentration on the order of 1,000,000 electrons cm^{-3}, appears above 300 km and is due to the entire ultraviolet spectrum. In fact, this peak is

associated with the effect of the diffusion of the ions. That is why below the F_2 region of the ionosphere the physical conditions are represented by the conditions of photoionization. Above there are conditions of diffusion leading to a gradual increase with altitude of the effect of the light ions (He^+, H^+) with respect to the heavy ions (N_2^+, O_2^+, NO^+, O^+). Thus, the terrestrial atmosphere is terminated by the metasphere indicating that the neutral elements (He, H) exist up to the highest altitudes, and by the protosphere signifying that at the limit there are hydrogen ions. Finally, it should be emphasized that the term magnetosphere is used to designate the extreme region where the effect of the magnetic field is of prime importance on the particles whose speeds are superthermal and do not follow from the distributions in the field of gravity.

3. Molecular Dissociation

The transition from the homosphere to the heterosphere is marked by molecular dissociation. Two molecules, O_2 and N_2, are influenced by solar radiation which can dissociate and ionize them.

The dissociation of molecular nitrogen ($N_2 \rightarrow N + N$) results directly from a predissociation of the molecule and ionization followed by ionic reactions which lead to nitrogen atoms. However, these atoms have a very short life time because they recombine by the following double process [8]

$$N + O_2 \rightarrow O + NO \tag{11}$$

and $\quad N + NO \rightarrow O + N_2 \tag{12}$

The result is that the molecule of nitrogen is practically not dissociated and that the atom of nitrogen is a minor constituent in the thermosphere.

The molecule of oxygen subjected to solar radiation is easily dissociated at altitudes below 100 km. There is no rapid process of reformation for molecular oxygen because the recombination of the atoms of oxygen depend essentially upon a three body process:

$$O + O + M \rightarrow O_2 + M \tag{13}$$

where M is the third body whose concentration decreases rapidly with altitude. In taking account of the pressure at 100 km, we see immediately that the life time of an atom of oxygen is several months.

Figure 1 shows the form in which the molecular dissociation must be viewed from one side of the mesopause to the other. In the mesosphere the conditions of photoequilibrium, where the number of photodissociations is compensated for by the number of recombinations, represent the daytime conditions. After sunset the concentration of atomic oxygen decreases because of the rapid recombination. In the thermosphere the time necessary to attain photodissociation equilibrium becomes very long and the transport by diffusion plays an important role. Thus, the increase with altitude of the atomic oxygen
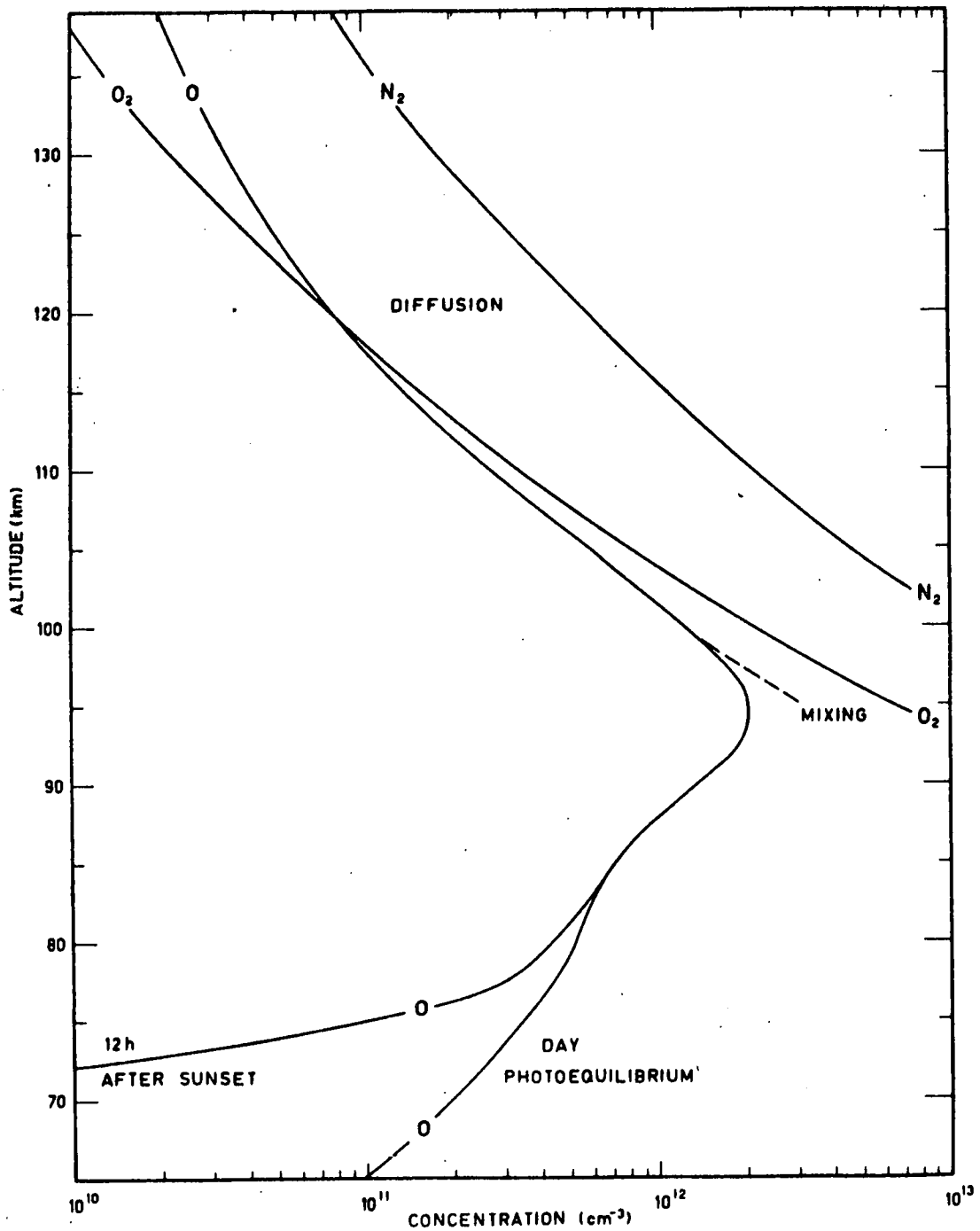
DISSOCIATION OF MOLECULAR OXYGEN BELOW AND ABOVE THE MESOPAUSE

FIGURE 1

concentration is limited and a concentration peak is reached between 90 and 100 km. Starting from this peak the molecular oxygen (molecular mass = 32) is subject to diffusion conditions and decreases more rapidly than the atomic oxygen (molecular mass = 16). The vertical distribution of molecular nitrogen remains practically the same because the molecular mass of the air mixture corresponds to M = 29 and that of N_2 is M = 28.

As a consequence, one obtains the following conditions above the mesopause. The vertical distribution of atomic oxygen produced by photodissociation depends essentially on mixing and diffusion processes. Some variations may occur because of major atmospheric movements. Advection can cause some changes if differences of concentration exist at different latitudes and longitudes. Dynamic transports can change the structure and vertical distribution while recombination is taking place. In any case, diffusion plays an important role. It transports upward the O_2 molecules which will replace those which are dissociated by the solar radiation or in the course of chemical reactions with the ions. Also the diffusion process transports downward the oxygen atoms which are formed above 100 km in order to allow a recombination in the lower thermosphere.

Because of the several vertical transports, a permanent equilibrium in the lower thermosphere cannot exist. Variations of concentration appear and the ratio $n(O_2)/n(O)$

at a certain altitude undergoes fluctuations. However, one
can fix the conditions at the lower limits of the zone of
diffusion by adopting average conditions. Table I presents
the value of the concentrations of the mesopause up to 150
km where the diffusion is complete. One sees that the ratio
$n(O_2)/n(O)$ is greater than unity below 110 km and that the
change to $n(O)/n(O_2) > 1$ occurs at an altitude (columns 4 and
5 of Table I) which depends upon the beginning of the diffu-
sion equilibrium of atomic oxygen. Molecular nitrogen re-
mains the principal element to 150 km.

These general conclusions must exist whatever may be
the atmospheric conditions. However, it is proper to insist
upon the sensitivity of the ratios $n(O_2)/n(O)$, $n(O)/n(N_2)$
as a function of the beginning of the diffusion equilibrium
level. It depends, in effect, on the atmospheric conditions
bound to the turbopause, that is to the region where the
atmospheric motions leading to mixing dominate over diffu-
sion.

TABLE I.- : Concentrations (cm^{-3}) of N_2, O_2 and O from the Mesopause to 150 km.

Altitude (km)	Temperature ($^{\circ}K$)	$n(N_2)$ (cm^{-3})	$n(O_2)$ (cm^{-3})	$n(O)^{(**)}$ (cm^{-3})	$n(O)^{(***)}$ (cm^{-3})
85	190	$1.5x10^{14}$	$3.9x10^{13}$	$7.2x10^{11}$	
90	197	$6.0x10^{13}$	$1.6x10^{13}$	$1.2x10^{12}$	
95	203	$2.5x10^{13}$	$6.7x10^{12}$	$2.0x10^{12}$	
100	210	$1.1x10^{13}$	$2.2x10^{12}$	$1.4x10^{12}$	
105	236	$4.4x10^{12}$	$8.1x10^{11}$	$5.9x10^{11}$	
110	262	$2.1x10^{12}$	$3.4x10^{11}$	$2.7x10^{11}$	$2.7x10^{11}$
115	293	$1.2x10^{12}$	$1.6x10^{11}$	$1.4x10^{11}$	$1.8x10^{11}$
120	324	$5.8x10^{11}$	$7.8x10^{10}$	$7.6x10^{10}$	$1.2x10^{11}$
130(*)	447	$1.8x10^{11}$	$2.2x10^{10}$	$3.4x10^{10}$	$5.4x10^{10}$
140(*)	570	$7.6x10^{10}$	$8.4x10^9$	$1.9x10^{10}$	$2.9x10^{10}$
150(*)	690	$3.8x10^{10}$	$3.9x10^9$	$1.1x10^{10}$	$1.8x10^{10}$

(*)   Above 120 km, the effect of the variation of temperature plays an important role in the ratios of the concentrations. The example chosen in this Table corresponds to model 4 of Figure 2.

(**) Diffusion equilibrium starts at 120 km.

(***) Diffusion equilibrium starts at 110 km.

4. Diffusion

In introducing the conditions of diffusion equilibrium below 120 km, we have seen how the concentration of atomic oxygen overtakes that of molecular oxygen and increases in comparison with that of molecular nitrogen. If the beginning of the level of diffusion were displaced to a higher level, for example from 120 to 150 km, the conditions would be different at the high altitudes. Thus, calculations made for atmospheres where the temperatures in the thermopause are between $2100^{\circ}K$ and $925^{\circ}K$ permit the presentation of the variation of the density as a function of altitude. We see in Figure 2 the double effect of temperature and diffusion. A factor of the order of 100 appears at 600 km. The effect of the beginning of diffusion (120 or 150 km) is important, because it can cause a variation of 100% in the density at 700 km. This is why it is fitting to consider the form in which the phenomenon of diffusion occurs.

Using the general formulas for diffusion derived by Chapman and Cowling[9], we can write the diffusion flow, F, for a minor constituent of mass m_1 in a gas of mass m in the following form:

$$F = \frac{3}{8} \frac{g_a \sqrt{\pi/2}}{\pi\sigma^2} (a/r)^2 (1 + m/m_1) (m/k)^{1/2} \frac{1}{T^{1/2}} \frac{n_1}{n} \times$$

$$\left[\left(X - \frac{m_1}{m} \right) + \left(\beta - \frac{2H}{r} \right) \left(X - 1 - \alpha_T \right) \right] \qquad (14)$$
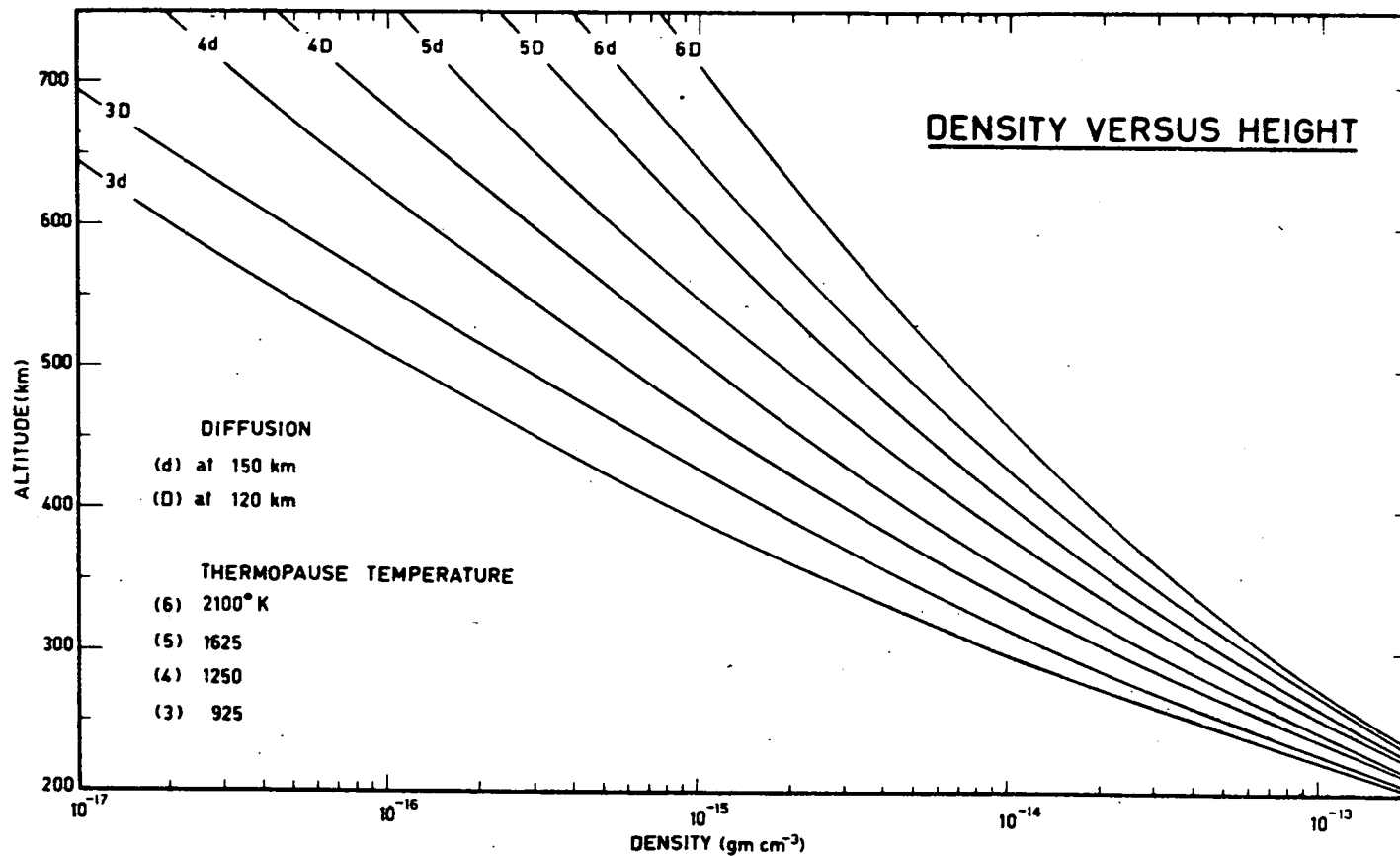
VERTICAL DISTRIBUTION OF DENSITY ACCORDING TO VARIOUS TEMPERATURES
REACHED AT THE THERMOPAUSE FOR DIFFUSION LEVELS AT 120 KM. AND 150 KM.;
RESPECTIVELY.

FIGURE 2

where n_1 and n are the concentrations of the constituents of mass m_1 and m. The diffusion is characterized by the effective cross-section $\pi\sigma^2$ where σ is the collision radius and by a thermal diffusion factor α_T. The atmosphere is characterized by the temperature T, the scale height H, and its gradient β, and by acceleration of gravity g and g_a for the distances to the center of the earth r and a.

The parameter X is the parameter of the vertical distribution of the constituent of mass m_1 and of concentration n_1. It is defined by

$$\frac{1}{n_1} \frac{\partial n_1}{\partial r} = X \frac{1}{n} \frac{\partial n}{\partial r} . \tag{15}$$

If X = 1, we obtain the conditions for mixing. If the value of X cancels the bracket of (14), the conditions for diffusion equilibrium are realized. Equation (14) indicates then the number of molecules or atoms that diffusion can transport. Vertical transport will be towards higher or lower altitudes according to the values of X and the ratio m_1/m which are the most important terms in the bracket of (14).

If we consider the conditions in the homosphere, X = 1, then (14) can be written (r_0 = radius of the earth),

$$F = 1.79 \times 10^{13} \left(\frac{r_0}{r}\right)^2 \left(1 + \frac{m}{m_1}\right)^{1/2} \left(\frac{m}{T}\right)^{1/2} \frac{n_1}{n} \left[1 - \frac{m_1}{m} - \alpha_T\left(\beta - \frac{2H}{r}\right)\right].$$

$$\tag{16}$$

We see that diffusion can transport almost the same number of particles to all the altitudes in the homosphere from the level of the earth up to the lower thermosphere. The variations are due to the very small decrease of $(r_o/r)^2$, to the difference in temperature $T^{1/2}$, and to the gradient of temperature when there is a coefficient of thermal diffusion. The application of (16) to the minor constituents such as argon, helium and atomic hydrogen at 100 km where we have the concentrations of Table I permit us to write:

$$F_{100 \text{ km}} \left(A^{40} \right) = -3.6 \times 10^{10} \text{ cm}^{-2}\text{sec}^{-1} \tag{17}$$

$$F_{100 \text{ km}} \left(He^{4} \right) = +8.8 \times 10^{7} \text{ cm}^{-2}\text{sec}^{-1} \tag{18}$$

$$F_{100 \text{ km}} (H) = +2.5 \times 10^{7} \text{ cm}^{-2}\text{sec}^{-1} \tag{19}$$

if $n(H) = 10^7 \text{ cm}^{-3}$.

If these flows do not differ much between 100 km and 120 km (only T varying), the diffusion times τ_D are very different. We have, in effect, the following relative values:

Altitude (km)	120	115	110	105	100
τ_D	1	1.5	2.5	4.5	9

We thus see how, from 100 km, the diffusion time decreases by about a factor of 10 in an altitude interval of only 20 km, while the mixing time must increase towards the

higher altitudes. From this point we very rapidly reach that region in which it is impossible to have an exchange of mass sufficiently great to counterbalance the effect of the diffusion. That is why we meet again in the observations (Townsend and Meadows [10], Johnson [11], Pokhunkov [12]) the variation of the argon/molecular nitrogen ratio at 110 ± 5 km.

The application of equations (14) and (15) indicates, then, that we pass from the state of equilibrium of mixing (X = 1) to the state of equilibrium of diffusion when the bracket of (14) is zero. In other words, the conditions of equilibrium of diffusion are written as:

$$\frac{\partial n_1}{\partial r} + \frac{n_1}{H_1} \left[1 + \left(\beta_1 - \frac{2H_1}{r} \right) \left(1 + \alpha_T \right) \right] = 0 \qquad (20)$$

for each constituent because F = 0.

If an escape flow exists in the exosphere, it is necessary to take this into account. The result is that, in the steady state case, it is necessary to have a balance between an atmospheric flow F and the escape F_C. We then write for an isothermal atmosphere:

$$F_C = F(r_0/r)^2 . \qquad (21)$$

The maximum diffusion flow $F_{D,Max}$ is fixed by (16) while the escape flow F_C at the critical level r_C is expressed for a constituent of mass m_1 by

$$F_C = n_{1,C}(g_C/2\pi)^{1/2} \, r_C \, H_{1C}^{-1/2} \, (1 + H_{1C}/r_C) \, e^{-r_C/H_{1C}} \qquad (22)$$

where $H_{1C} = kT/m_1 g$. The flow F_C depends, in major part, on the ratio r_C/H_{1C} which enters in the exponential of (22).

The comparison of the two flows $F_{D, \, Max}$ and F_C permit us to draw immediate conclusions [13, 14] on the behavior of the vertical distribution of the constituents. To this end, it is a question of first introducing the value of the concentration n_{1C} into (22) for a selected temperature of the thermopause. If we introduce the concentration that is obtained by integration of (20), that is by integrating the expression giving the conditions for the equilibrium of diffusion, we obtain the maximum value of the escape flow, $F_{C, \, Max}$. If

$$F_{D \, Max} > F_{C \, Max} \, , \qquad (23)$$

the conditions of equilibrium of diffusion can be applied since the diffusion flow always supports the escape flow. This is the case for helium where the diffusion flow at the start of the heterosphere is always greater than the flow resulting from its escape into interplanetary space. In other words, the flow $F_{100 \, km}(He^4)$ given by (18) is always greater than the flow of He^4 atoms which can result from its generation in the earth's crust ($< 10^7 \, He^4 \, cm^{-2} sec^{-1}$). As a consequence, the steady state case for He^4 is represented in the region of diffusion by a vertical distribution of

diffusion equilibrium defined by (20). If

$$F_{D\ Max} < F_{C\ Max} \quad , \tag{24}$$

the result is that the diffusion flow cannot support the escape flow. In this case, the concentration n_{1C} decreases to the point where it is adapted to the equality $F_C = F_{C\ Max}$. There will then be, as a function of altitude, a decrease in the thermospheric concentration of the constituent, in order to account for the maximum value of F_D. This is the case for atomic hydrogen when (24) is applied. In fact, the equation

$$F_C(H) = F_{D\ Max}\ (r_o/r)^2 \quad , \tag{25}$$

when written for atomic hydrogen H, must be applied to determine the vertical distribution of the hydrogen.

The application of (18) and (19) and of (23) and (25) leads to the following order of magnitudes of concentrations at 500 km:

$T(^{\circ}K)$	2000	1400	1200	1000
$n_{500}(He^4), (cm^{-3})$	2.0×10^6	1.9×10^6	1.8×10^6	1.6×10^6
$n_{500}(H)\ , (cm^{-3})$	1×10^3	5×10^3	1×10^4	3×10^4

We see, then, that the behavior of helium and of hydrogen is different. Helium is in diffusion equilibrium and at practically the same concentration at 500 km. This depends on the value of the concentration taken at the beginning of

the diffusion equilibrium level. Hydrogen, on the other hand, cannot attain diffusion equilibrium conditions in the thermosphere. Its concentration at 500 km decreases when the escape velocity, v_C, increases with temperature. We have, according to (22) and (25),

$$F_{D\ Max} = F_C = n_{1C}\ v_{1C} \ ; \qquad (26)$$

that is, a decrease of n_{1C} when v_{1C} increases. The increase of v_{1C} is very rapid as T increases, and that is why the concentration of hydrogen in the exosphere varies by more than a factor of 10 when the temperature varies from $2000°K$ to $1000°K$.

5. Heating by Solar Ultraviolet Radiation

When solar radiation is absorbed, we can consider that, in a general form, the production of heat P takes the form

$$P = nK\ E_{uv}\ e^{-\tau}\ d\tau \qquad (27)$$

where K is the appropriate cross-section of absorption , E_{uv} the available solar energy for the heating, and τ the optical thickness. Integration of (27) gives the flux of heat, E, at an altitude z,

$$E = E_\infty + E_{uv}\ (1 - e^{-\tau}) \qquad (28)$$

where E_∞ is the energy of a source of heat external to the

atmospheric layer considered.

The atmosphere is heated by all the radiations of the solar spectrum. But the radiations which are most important for the heating of the thermosphere belong principally to the spectral region between 900Å and 200Å. The energy that the thermosphere receives ($E \geq 1$ erg $cm^{-2}sec^{-1}$) in twelve hours is equivalent to the total kinetic energy of a thermospheric column. Thus, there exists in the thermosphere a region, centered appropriately around 150 km, where the solar flux absorbed is actually effective in the heating. At 100 km and in the lower regions the solar flux absorbed is also very important. But the kinetic energy of an atmospheric column is very high compared with the flux received during the course of a day. This produces, then, only a small increase of the temperature. At the very high altitudes where the optical thickness becomes small

$$E = \tau \, E_{uv} \tag{29}$$

and the ultraviolet heating is very weak.

It is necessary to introduce a system of cooling which does not permit an excessive heat accumulation in the interior of the thermosphere. The three modes of transport of heat convection, radiation, and conduction can be introduced. Convection plays scarcely any role at the higher altitudes of the thermosphere. The loss by radiation is limited to the oxygen atom at 65 μ and does not exist for the molecules

N_2 and O_2. The transport by conduction is the most important process because it depends on the gradient of the temperature. It is, thus, this phenomenon that must be considered first for the loss of heat in the thermosphere.

6. The Conduction of Heat

The density of the heat flow, E_c, is expressed by

$$E_c = - \lambda_c \text{ grad } T, \tag{30}$$

and the equation of continuity is written

$$\rho \, c_v \, \frac{\partial T}{\partial t} + \text{div. } E_c = P - L. \tag{31}$$

The coefficient λ_c is the thermal conductivity which we can write for each constituent of the atmosphere in the general form

$$\lambda_c = A \, T^{2/3} \tag{32}$$

where A is an appropriate constant.

The coefficient c_v is the specific heat at constant volume, while P and L are the production and the loss of heat per unit time and volume; respectively.

By writing

$$\theta = \int_{T_2}^{T} \frac{\lambda}{\lambda_2} \, dT, \tag{33}$$

where T_2 is the temperature at the altitude where we introduce the boundary conditions, we obtain for (31) the expression

$$\frac{\partial \theta}{\partial t} = \frac{A_1 T^{2/3}}{n} \left[\nabla^2 \theta + \frac{P - L}{A T_2^{2/3}} \right] \tag{34}$$

where A_1 is also a constant.

When we study the cooling of an atmospheric layer where the conditions are simple, the application of (34) shows that the time of cooling is inversely proportional to the concentration and to the square of the distance. Thus, the cooling time is about 100 times more rapid at 500 km than at 200 km. The practical results from the point of view of the behavior of the atmosphere are that: (1) due to the decrease of E with altitude, equation (29), and (2) due to the decrease of the cooling time with altitude, the tendency to a vertical isothermy is the essential characteristic of the upper atmosphere. In fact, the higher regions of the atmosphere ($>$ 500 km) can thus be considered as being best represented by a vertical isothermy and by a state of diffusion equilibrium.

Finally, if we assume steady state conditions, we see that the vertical distribution of the temperature depends on the equality of the effects of the conduction and heating flows. We have then for temperatures between 400°K and 1600°K in a molecular atmosphere the following values of the temperature gradient

$$(dT/dz)_{km} = (20 \pm 6)\ E \tag{35}$$

if E is expressed in erg $cm^{-2}sec^{-1}$, or for the gradient β of the scale height

$$\beta = (0.46 \pm 0.10)E . \tag{36}$$

This elementary examination of the process of conduction justifies the existence of the gradients required in the thermosphere in order to explain the high densities deduced from the variation of the periods of artificial satellites. Moreover, the diurnal variation of the density is explained by the double fundamental processes of ultraviolet heating and conduction leading to a diurnal variation of the temperature. Finally, the ultraviolet heating is variable from the fact that the ultraviolet flux changes with solar activity. The density of the atmosphere whose observed variation is correlated with that of a solar activity index is explained by a variation of temperature.

7. The Variation of the Constitution of the Upper Atmosphere as a Function of the Temperature

When the conditions at the onset of diffusion are fixed in the thermosphere, the vertical distribution of the constituents can be determined throughout the upper atmosphere. In effect, the vertical distribution of the temperature determines that of the pressure, of the density, and of the concentration of each constituent. In the steady state case, the diffusion equilibrium is applicable and the differences between the diverse vertical distributions of the total density depend upon the conditions of temperature which affect the several constituents differently.

It is, then, necessary to have an idea of the ratios of the concentrations of the principal constituents in the

lower atmosphere where we must introduce the diffusion conditions. First, it is necessary to account for the dissociation of the oxygen (cf. Fig. 1) and the onset of diffusion (cf. Table 1). It is not possible to represent by a single atmospheric model all the conditions of the lower thermosphere. Some seasonal variations must appear following the position of the sun and, consequently, a latitude effect must exist. Nevertheless, the essential condition to be fixed at the beginning of the diffusion level is the ratio of the principal constituents. We have emphasized above that molecular nitrogen must be the principal constituent of the lower thermosphere and that molecular oxygen must be more abundant than atomic oxygen below the zone of equilibrium of diffusion.

It is fitting, however, to note here that this last concept is not accepted in certain recent works. Hinteregger [15] and Norton, Van Zandt and Denison [16], consider that atomic oxygen very rapidly becomes the most important element. For example, they indicate that $n(O)/n(N_2) \geq 1$ at 130 km and that $n(O)/n(O_2) > 1$ at 100 km. Such considerations are hardly acceptable if we refer to the results obtained by the mass spectrographic measurements in the USA [17] and in the USSR [18]. These experimental results approach much more closely the conditions adopted by Nicolet [19]. For example, the molecular mass at 200 km takes on the following values:

Nicolet [19]   Townsend [17]   Pokhunkov [18]   Van Zandt [16]

24.3 ± 0.5   > 22   24.4   18.7

Hinteregger [15]

   ≤ 18.5

It is then difficult to avoid the conclusion that the mole-
cular mass at 200 km is greater than 22, that is

$$n(N_2) \geq n(0) \text{ at 200 km } ,$$

and thus the adoption of values of $M < 20$ cannot be justified
in actual circumstances.

By adopting as initial conditions those which are given
in Table I at 120 km, we can choose different temperature
gradients and thereby obtain the vertical distributions of the
concentrations of N_2, 0 and O_2. The several vertical distri-
butions must be adapted to equation (2) and the ensemble to
equations (28) and (34).

A first result applied at 150 km (Table II), shows
the small variation of density at this altitude. A variation
of temperature greater than $100^\circ K$ gives only a small variation
of the density $\rho = (2.5 \pm 0.1) 10^{-12} \text{ gm cm}^{-3}$. But at 400 km,
for example, the result is such that (Table III) the density
varies by more than a factor of 5. From 500 km, the condi-
tions become different because of the presence of helium [20].
It is possible to account for the presence of helium in
hydrostatic equilibrium up to 3000 km. As for hydrogen, it

TABLE II.- : Physical Parameters at 150 km.

No	$T^{(1)}$	$H^{(2)}$	$\beta^{(3)}_{150-160}$	$H_\rho^{(4)}$	$\rho^{(5)}$	$p^{(6)}$	$M^{(7)}$	$n^{(8)}$
0	877	29.8	0.67	20.0	2.41(-12)	5.0(-6)	26.2	5.6(+10)
1	873	29.6	0.56	20.7	2.43	5.0	26.2	5.6
1.5	863	29.3	0.53	20.9	2.45	5.0	26.2	5.6
2	852	28.9	0.48	21.0	2.48	5.0	26.2	5.7
3	829	28.1	0.44	21.0	2.54	5.0	26.1	5.8
4	803	27.3	0.38	21.0	2.59	5.0	26.1	6.0
5	777	26.4	0.35	21.1	2.62	4.9	26.1	6.0
6	751	25.6	0.31	20.8	2.65	4.8	26.1	6.1
7	726	24.7	0.28	20.4	2.61(-12)	4.5(-6)	26.1	6.0(+10)

(1) temperature in $^\circ K$;

(2) Atmospheric scale height in km;

(3) Gradient of H for the height interval indicated in the table;

(4) Density scale height parameter;

(5) Density in gm. cm^{-3} with (-12) for 10^{-12};

(6) Pressure in mm Hg with (-6) for 10^{-6};

(7) Mean molecular mass with the mass of atomic oxygen M = 16;

(8) Total concentration in cm^{-3} with (+10) for 10^{10},

TABLE III.- : Atmospheric conditions at 400 km.

No	$T^{(1)}$	$H^{(2)}$	$\beta^{(3)}_{400-420}$	$H_\rho^{(4)}$	$\rho^{(5)}$	$P^{(6)}$	$M^{(7)}$	$n^{(8)}$
0	2086	96.9	0.14	91.2	1.93(-14)	1.2(-7)	20.6	5.6(+8)
1	1826	87.3	0.13	82.3	1.53	8.7(-8)	20.0	4.6
1.5	1707	82.8	0.12	74.6	1.36	7.3	19.7	4.1
2	1597	78.6	0.11	73.6	1.19	6.1	19.4	3.7
3	1412	71.6	0.10	67.7	9.11(-15)	4.2	18.9	2.9
4	1272	66.2	0.09	62.7	6.89	3.0	18.4	2.2
5	1155	61.5	0.08	59.2	5.16	2.1	18.0	1.7
6	1059	57.6	0.07	55.4	3.86	1.4(-8)	17.6	1.3(+8)
7	978	54.2	0.08	52.6	2.81	9.9(-9)	17.3	9.8(+7)

same explanations as for TABLE II.

is necessary to account for the fact that its vertical distribution in the thermosphere depends upon the diffusion flow. Figure 3 indicates clearly that below 1120 km its vertical distribution is not far from that of an atmosphere where mixing takes place. It is necessary to reach altitudes on the order of 500 km in order to be able to represent the vertical distribution of hydrogen by a state of diffusion equilibrium. We consequently reach the conclusion that the concentrations of hydrogen and of helium have a different behavior at 500 km. If helium is characterized by a concentration which is almost constant, that of hydrogen varies as a strong function of temperature. Figure 4 is an example of the conditions existing at 500 km and shows how the differences appear up to 3000 km.

In regrouping the set of conditions which are attached to O, N_2, He and H, we obtain the numerical results [14] presented in Table IV. We realize immediately the importance of helium when the temperatures are high. It is appropriate to indicate that these results can differ from those published by other authors (Harris and Priester, Paetzold and Kallman-Bijl, 1962[*]). This is easily explained because, when the thermal diffusion (α_T in equation 16) is neglected, the calculated concentrations of He and H as a function of altitude are not correct. Moreover, considerable errors are introduced when the diffusion equilibrium is applied to hydrogen in the thermosphere.

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[*] COSPAR meeting, May, 1962, being published.
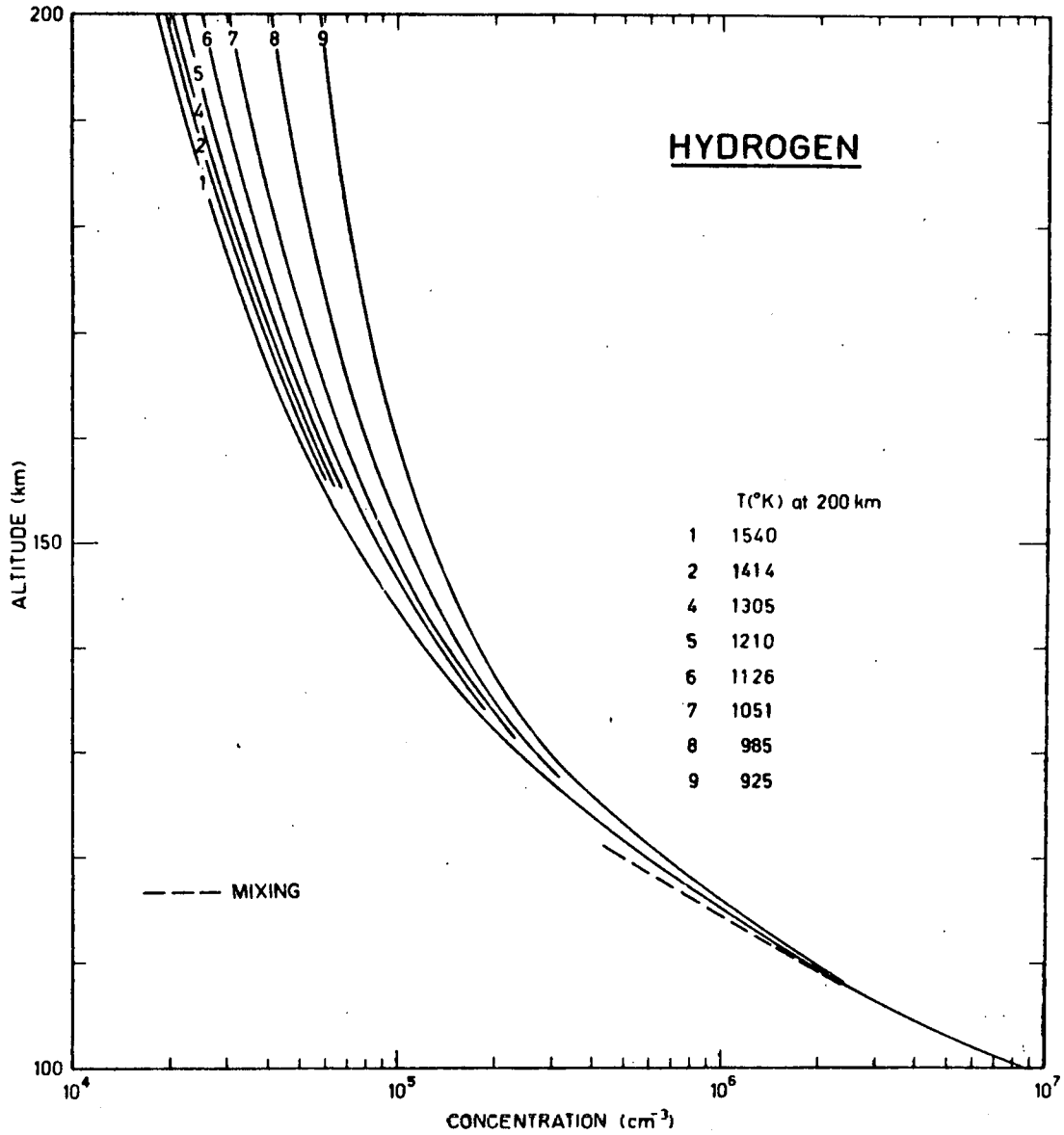
VERTICAL DISTRIBUTION OF ATOMIC HYDROGEN BETWEEN 100 KM AND 200 KM

(FROM KOCKARTS AND NICOLET, ANNALES GÉOPHYS., 18, 269, 1962)

FIGURE 3
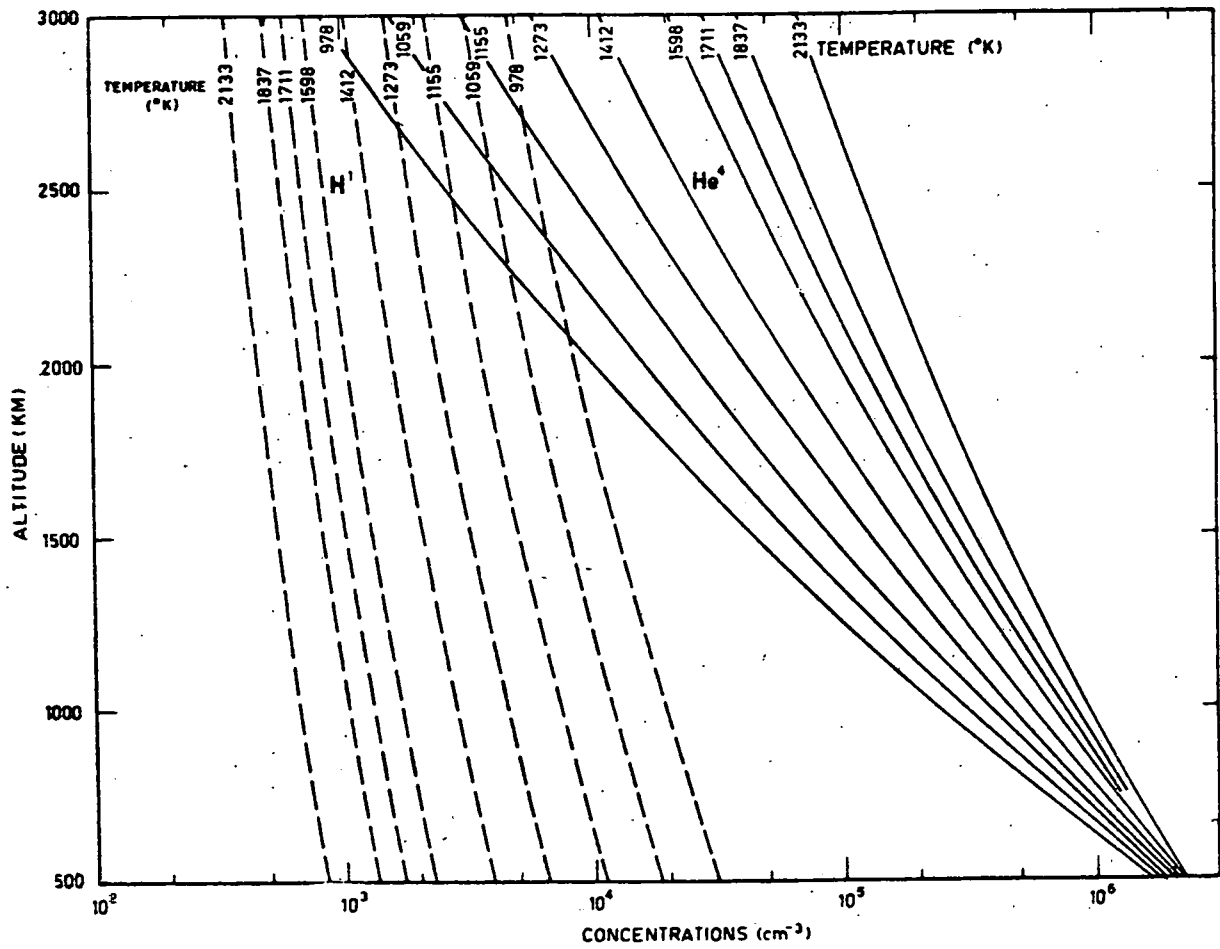
VERTICAL DISTRIBUTION OF HELIUM AND ATOMIC HYDROGEN FROM 500 KM UP TO 3000 KM IN DIFFUSION EQUILIBRIUM (FROM KOCKARTS AND NICOLET, ANNALES GÉOPHYS., 18, 269, 1962).

FIGURE 4

TABLE IV : CONCENTRATIONS (cm^{-3}), MASS (O=16) AND TOTAL DENSITY (gm cm^{-3}) FROM 500 km TO 3000 km FOR VARIOUS THERMOPAUSE TEMPERATURES.

Temperature (°K)	2133	1837	1711	1598	1412	1273	1155	1059	978
500 km									
n(O)	1.58×10^8	1.25×10^8	1.10×10^8	9.45×10^7	6.91×10^7	4.96×10^7	3.49×10^7	2.42×10^7	1.63×10^7
n(N$_2$)	4.85×10^7	2.90×10^7	2.17×10^7	1.59×10^7	8.39×10^6	4.35×10^6	2.18×10^6	1.08×10^6	5.17×10^5
n(He^4)	2.26×10^6	2.25×10^6	2.20×10^6	2.19×10^6	2.11×10^6	2.01×10^6	1.90×10^6	1.77×10^6	1.66×10^6
n(H)	8.27×10^2	1.34×10^3	1.74×10^3	2.28×10^3	3.91×10^3	6.51×10^3	1.10×10^4	1.85×10^4	3.11×10^4
Mass	18.9	18.2	17.9	17.6	17.0	16.5	16.1	15.7	15.3
Total Density	6.71×10^{-15}	4.8×10^{-15}	4.0×10^{-15}	3.3×10^{-15}	2.3×10^{-15}	1.5×10^{-15}	1.0×10^{-15}	7.1×10^{-16}	4.7×10^{-16}
750 km									
n(O)	2.45×10^7	1.49×10^7	1.11×10^7	8.17×10^6	4.32×10^6	2.30×10^6	1.18×10^6	6.00×10^5	2.98×10^5
n(N$_2$)	1.84×10^6	6.98×10^5	3.96×10^5	2.20×10^5	6.58×10^4	2.01×10^4	5.80×10^3	1.68×10^3	4.71×10^2
n(He^4)	1.42×10^6	1.32×10^6	1.24×10^6	1.19×10^6	1.06×10^6	9.33×10^5	8.15×10^5	7.04×10^5	6.08×10^5
n(H)	7.36×10^2	1.17×10^3	1.51×10^3	1.95×10^3	3.28×10^3	5.37×10^3	8.89×10^3	1.47×10^4	2.42×10^4
Mass	16.2	15.6	15.2	14.8	13.8	12.6	11.1	9.5	7.8
Total Density	7.5×10^{-16}	4.4×10^{-16}	3.2×10^{-16}	2.4×10^{-16}	1.2×10^{-16}	6.8×10^{-17}	3.7×10^{-17}	2.1×10^{-17}	1.2×10^{-17}
1000 km									
n(O)	6.18×10^6	2.05×10^6	1.32×10^6	8.34×10^5	3.27×10^5	1.31×10^5	4.99×10^4	1.92×10^4	7.15×10^3
n(N$_2$)	1.65×10^5	2.17×10^4	9.45×10^3	4.05×10^3	7.18×10^2	1.34×10^2	2.31×10^1	4.04×10^0	--
n(He^4)	9.26×10^5	8.03×10^5	7.29×10^5	6.71×10^5	5.53×10^5	4.56×10^5	3.70×10^5	2.97×10^5	2.39×10^5
n(H)	6.60×10^2	1.03×10^3	1.32×10^3	1.69×10^3	2.79×10^3	4.48×10^3	7.28×10^3	1.18×10^4	1.91×10^4
Mass	14.2	12.7	11.8	10.7	8.5	6.6	5.4	4.6	4.1
Total Density	1.3×10^{-16}	6.1×10^{-17}	4.0×10^{-17}	2.7×10^{-17}	1.2×10^{-17}	6.5×10^{-18}	3.8×10^{-18}	2.5×10^{-18}	1.8×10^{-18}
1250 km									
n(O)	8.97×10^5	3.21×10^5	1.80×10^5	9.85×10^4	2.93×10^4	9.00×10^3	2.61×10^3	7.67×10^2	2.20×10^2
n(N$_2$)	5.64×10^3	8.47×10^2	2.90×10^2	9.64×10^1	1.05×10^1	1.24×10^0	--	--	--
n(He^4)	6.21×10^5	5.05×10^5	4.43×10^5	3.93×10^5	3.03×10^5	2.33×10^5	1.77×10^5	1.33×10^5	1.00×10^5
n(H)	5.97×10^2	9.2×10^2	1.16×10^3	1.48×10^3	2.40×10^3	3.79×10^3	6.05×10^3	9.64×10^3	1.54×10^4
Mass	11.1	8.8	7.5	6.4	5.0	4.4	4.1	3.9	3.6
Total Density	2.8×10^{-17}	1.2×10^{-17}	7.7×10^{-18}	5.2×10^{-18}	2.8×10^{-18}	1.8×10^{-18}	1.3×10^{-18}	9.2×10^{-19}	7.0×10^{-19}

TABLE IV (CONT.)

1500 km

n(O)	2.01×10^5	5.65×10^4	2.79×10^4	1.33×10^4	3.06×10^3	7.34×10^2	1.65×10^2	3.76×10^1	8.43×10^0
$n(N_2)$	4.12×10^2	4.06×10^1	1.20×10^1	2.91×10^0	--	--	--	--	--
$n(He)$	4.27×10^5	3.27×10^5	2.78×10^5	2.39×10^5	1.72×10^5	1.25×10^5	8.86×10^4	6.25×10^4	4.43×10^4
n(H)	5.43×10^2	8.25×10^2	1.03×10^3	1.30×10^3	2.08×10^3	3.23×10^3	5.09×10^3	7.97×10^3	1.25×10^4
Mass	7.8	5.8	5.1	4.6	4.2	4.0	3.9	3.7	3.3
Total Density	8.2×10^{-18}	3.7×10^{-18}	2.6×10^{-18}	1.9×10^{-18}	1.2×10^{-18}	8.5×10^{-19}	6.0×10^{-19}	4.3×10^{-19}	3.2×10^{-19}

2000 km

n(O)	1.32×10^4	3.40×10^3	9.34×10^2	3.40×10^2	5.00×10^1	7.65×10^0	1.08×10^0	--	--
$n(N_2)$	--	--	--	--	--	--	--	--	--
$n(He)$	2.16×10^5	1.48×10^5	1.19×10^5	9.63×10^4	6.14×10^4	3.98×10^4	2.52×10^4	1.59×10^4	1.00×10^4
n(H)	4.57×10^2	6.76×10^2	8.35×10^2	1.04×10^3	1.60×10^3	2.43×10^3	3.71×10^3	5.64×10^3	8.60×10^3
Mass	4.7	4.2	4.1	4.0	3.9	3.8	3.6	3.2	2.6
Total Density	1.8×10^{-18}	1.1×10^{-18}	8.2×10^{-19}	6.5×10^{-19}	4.1×10^{-19}	2.7×10^{-19}	1.7×10^{-19}	1.1×10^{-19}	8.1×10^{-20}

2500 km

n(O)	1.29×10^3	1.45×10^2	4.60×10^1	1.38×10^1	1.30×10^0	--	--	--	--
$n(N_2)$	--	--	--	--	--	--	--	--	--
$n(He)$	1.18×10^5	7.35×10^4	5.59×10^4	4.29×10^4	2.46×10^4	1.45×10^4	8.24×10^3	4.69×10^3	2.69×10^3
n(H)	3.92×10^2	5.67×10^2	6.91×10^2	8.47×10^2	1.27×10^3	1.88×10^3	2.80×10^3	4.15×10^3	6.17×10^3
Mass	4.1	4.0	4.0	4.0	3.8	3.7	3.2	2.6	1.9
Total Density	8.2×10^{-19}	4.9×10^{-19}	3.7×10^{-19}	2.9×10^{-19}	1.7×10^{-19}	9.9×10^{-20}	5.9×10^{-20}	3.8×10^{-20}	2.8×10^{-20}

3000 km

n(O)	1.37×10^2	1.12×10^1	3.12×10^0	--	--	--	--	--	--
$n(N_2)$	--	--	--	--	--	--	--	--	--
$n(He)$	6.88×10^4	3.93×10^4	2.85×10^4	2.09×10^4	1.09×10^4	5.85×10^3	3.04×10^3	1.58×10^3	8.25×10^2
n(H)	3.42×10^2	4.84×10^2	5.83×10^2	7.06×10^2	1.04×10^3	1.50×10^3	2.18×10^3	3.16×10^3	4.59×10^3
Mass	4.0	4.0	3.9	3.9	3.7	3.4	2.8	2.0	1.5
Total Density	4.6×10^{-19}	2.6×10^{-19}	1.9×10^{-19}	1.4×10^{-19}	7.4×10^{-20}	4.1×10^{-20}	2.4×10^{-20}	1.6×10^{-20}	1.3×10^{-20}

8. The Solar Radiation and Variations of Temperature

In examining the variations of the density as functions of the temperature, (cf. Table IV), we must conclude that the variations of the density deduced from the variation of the periods of satellites are linked to the variations of temperature. From 1957 to 1962, many observations of satellites have been made. These demonstrate that the ultraviolet flux from the sun warming the thermosphere has decreased markedly from the maximum of solar activity till 1962. This result is generally explained by the introduction of a solar activity index. The radioelectric solar flux at 10 cm is usually chosen because the observations made at Ottawa cover more than one cycle of solar activity. It has likewise been shown that it is at this wavelength that the flux varies most strongly.

Nevertheless, it is best to be very careful in the study of the correlation between a radioelectric flux at a selected wavelength and the temperature of the thermopause. A monochromatic emission cannot account for the variations of the integrated effect of the solar ultraviolet radiation. A lack of correlation between the solar emission at 10 cm and the temperature of the thermopause in the course of a cycle of solar activity cannot be interpreted by a hypothesis such as that of corpuscular radiation. It is necessary, on the contrary, to account for the fact that the use of a single wavelength for an index of solar activity does not represent the complete set of conditions. Moreover, it is necessary to

consider that the conditions of the low thermosphere are variable in the course of the year and thereby modifies the boundary conditions of all the upper atmosphere.

The case of disturbances is different. When a magnetic storm occurs, it is certain that the upper atmosphere is influenced on a worldwide scale. That is why it is not the result of a direct effect of charged particles at 150 km and at the equator. It is necessary that the general warming of the atmosphere be due to a type of heating with a hydromagnetic character. This can appear in the disturbed conditions of the magnetic field.

9. The Constitution of the Ionosphere.

The constitution of the conventional ionosphere, limited at its summit by the maximum of concentration of the F_2 region, can be determined by the study of photoionization equilibrium. In other words, the ultraviolet radiation of the sun ionizes the principal constituents N_2, O_2 and O. Thus the ions N_2^+, O_2^+ and O^+ must appear. There are, however, two exceptions: cosmic radiation, as has been shown by Nicolet (cf. Nicolet-Aikin[7]), is responsible for the ionization in the lower part of the region D and must decrease from the equator to the poles with respect to the penetration of these particles. The Lyman-α chromospheric radiation does not ionize the oxygen and nitrogen but direct ionization of the NO molecule is possible. However, the mass-spectrometric observations (cf. Johnston[11], Istomin[21]) demonstrate that the NO^+ ion

is so abundant that it must result from atmospheric reactions and not from direct photoionization of NO in the E and F regions. Finally, the addition in the E region of ions such as Mg^+, Ca^+ (Istomin [22]) must result from the effect of meteors (Nicolet [23]).

The process of photoionization depends on the energy of the ultraviolet radiations penetrating the ionosphere. At present we see that the conditions for the study of photo-ionization, in spite of the observations made by sounding rockets, are very different, although the authors state that the ionospheric structure is explained. For example, Ivanov-Kholodny [24] uses a solar ultraviolet flux 10 times greater than that presented by Watanabe and Hinteregger [25]. If such differences are introduced for the study of the effects of ionization, it is certain that the differences are even greater in the application of ionospheric reactions. This results first from the fact that the experimental measure-ments of the reaction rate coefficients are not sufficiently numerous and are yet too crude.

Also, the system of the reactions which can occur is extremely complex. Figure 5 indicates [26] how the diverse reactions appear after the production of ions under the effect of solar radiation. We establish first that only the NO^+ ion disappears solely by dissociative recombination with an electron. This is due to the fact that the dissociation energy of NO^+ is the highest; 10.88 eV. This ion can then

I O N O S P H E R I C   R E A C T I O N S

NO + hν — I(NO) → e + NO$^+$ — α(NO) → N + O

γ_6 γ_7

N N_2
+ $+^2$

O$_2$ + hν — I(O$_2$) → e + O$_2^+$ — α(O$_2$) → O + O

γ_9 $\gamma_{2(5)}$ γ_4

O$_2$ NO
$+^2$ +

O + hν — I(O) → e + O$^+$ —— + N$_2$ — γ_1 → N + NO$^+$

                          —— + NO — γ_3 → O + NO$^+$

γ_8

O
+

O$_2$
$+^2$

N$_2$ + hν — I(N$_2$) → e + N$_2^+$ — α(N$_2$) → N + N

                          + O — γ_{14}^{**}→ N + NO$^+$ (^{3}Π)

γ_{11} + O$_2$ — γ_{15}^{*}→ NO + NO$^+$ (^{3}Π)

+   NO
N   +

γ_{10}

N + hν — I(N) → e + N$^+$ —— + O$_2$ — γ_{12}→ O + NO$^+$ (^{3}Π)

                          —— + NO — γ_{13}→ O + NO$^+$ (^{3}Π)
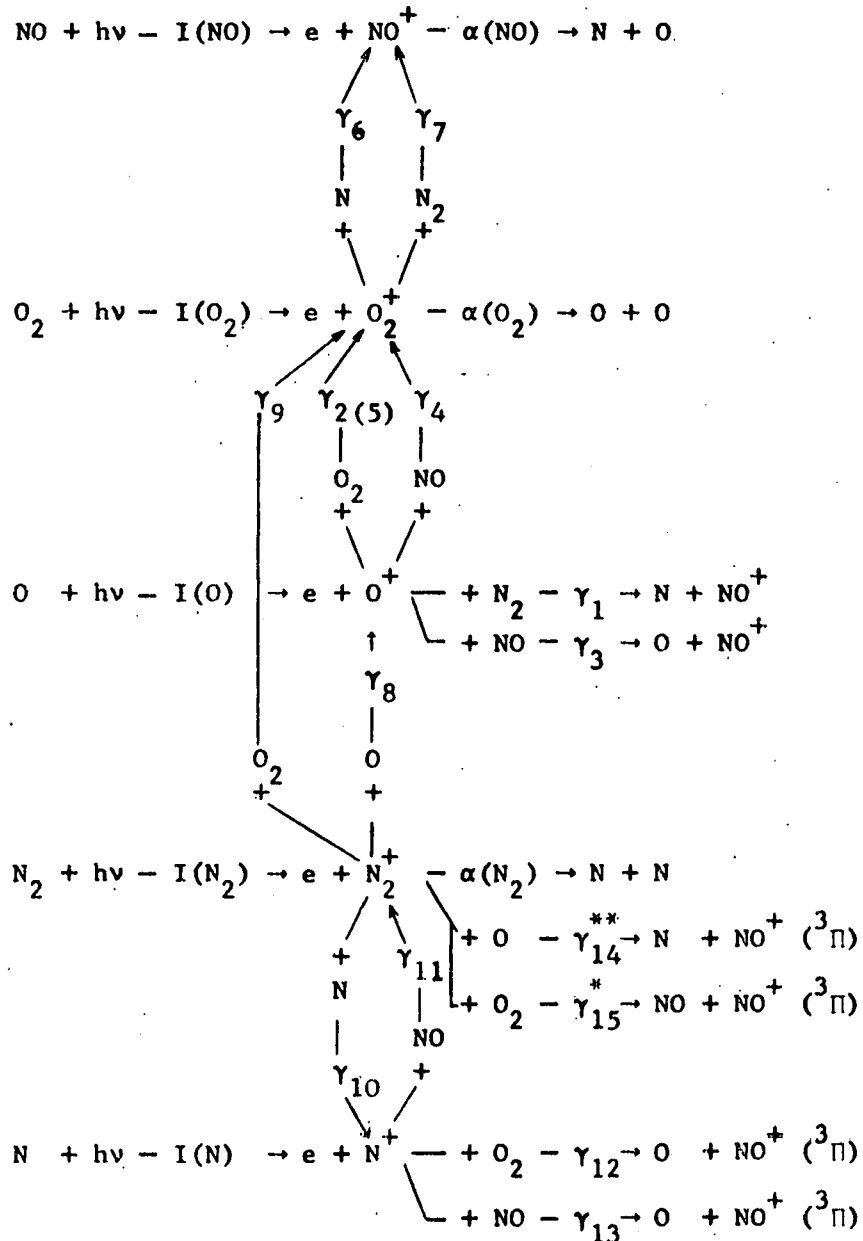
IONOSPHERIC REACTIONS. PHOTOIONIZATION WITH RATE
COEFFICIENT I, DISSOCIATIVE RECOMBINATION WITH COEF-
FICIENT α, CHARGE TRANSFER OR ION-ATOM INTERCHANGE
WITH RATE COEFFICIENT ૪.

FIGURE 5

exist in all circumstances for longer times than the other ions. Also, we see that N_2^+ undergoes, in addition to dissociative recombination, a charge transfer with oxygen (O and O_2). The result is that this ion is less abundant at the lower altitudes than in the F region because of the high concentrations of O_2 and O. Thus, the O^+ ion results completely from the direct photoionization of atomic oxygen and partially from the photoionization of nitrogen. The disappearance of O^+ is caused simultaneously by ion-atom interchange with O_2 and N_2. This ion-atom interchange process leads to O_2^+ and NO^+ We see that it is necessary to know the values of the reaction coefficients very precisely in order to establish the ionospheric recombination. Finally, Figure 5 shows that O_2^+ can lead to NO^+ by a reaction with N_2. Even if such a reaction has a low probability, it cannot be neglected in the D region where molecular nitrogen is very abundant.

From the above we can state, in the first approximation, that the following reactions will be important:

Charge transfer processes

$$(\gamma_8) \; ; \; N_2^+ + O \rightarrow O^+ + N_2 \qquad\qquad (37a)$$

$$(\gamma_9) \; ; \; N_2^+ + O_2 \rightarrow O_2^+ + N_2 \qquad\qquad (37b)$$

Ion-atom interchange reactions

$$(\gamma_1) \; ; \; O^+ + N_2 \rightarrow NO^+ + N \qquad\qquad (38a)$$

$$(\gamma_2) \; ; \; O^+ + O_2 \rightarrow O_2^+ + O \qquad\qquad (38b)$$

Thus, at all altitudes

$$(\gamma_{12}) \; ; \; N^+ + O_2 \rightarrow O + NO^+ \tag{39}$$
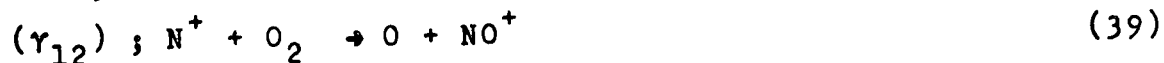
and finally, in the lower ionosphere,

$$(\gamma_7) \; ; \; O_2^+ + N_2 \rightarrow NO^+ + NO \tag{40}$$
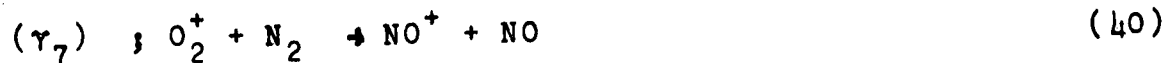
A full description of the structure of the ionosphere requires then many more experimental studies and it is also necessary to extend the domain of the studies with the aid of sounding rockets.

When the effect of the ionospheric reactions decreases because of the concentration decrease, it is necessary to introduce the effect of diffusion of ions and electrons. We can say that the F_2 peak results from the effect of the decrease of the electronic recombination and the augmentation of the diffusion effect with altitude. This is why it is necessary to study the upper ionosphere in the frame of diffusion. Thus, when Nicolet [20] showed that the existence of neutral helium also required the presence of ionized helium, the observations (Hanson [27], Bourdeau et al. [28], Bauer and Jackson [29], etc) indicated the presence of this ion above 500 km. Further, Mange [30,31] had shown how minor ions such as H^+ and He^+ could appear with concentrations increasing with altitude. It then required a new study [32] of the problem of the equilibrium of diffusion of the ions and electrons.

When an ionized gas is neutral, the condition of neutrality is

$$n_e = \Sigma \; Z_i \; n_i^+ \tag{41}$$

and

$$\frac{dn_e}{dr} = \frac{d(\Sigma \ Z_i \ n_i^+)}{dr} \qquad (42)$$

where n_e is the concentration of electrons of mass m_e and charge $-e$, n_i^+ is the concentration of ions of mass m_i and charge $+ Z_i e$.

If the ionized gas is in a steady state in the presence of external forces which derive from potential functions, the velocities have a Maxwellian distribution (Chapman and Cowling[9]). The presence of a magnetic field does not affect the concentration distribution. However, when the atmosphere rotates about the axis of z with angular velocity ω_e, the vertical distribution depends on the field of the centrifugal force and also on the magnetic field. The potential at a distance r from the earth's center is

$$\Phi = \varphi + \omega_e^2 \ (x^2 + y^2) - \frac{e}{m} \ \omega_e \ (x^2 + y^2) \ A \qquad (43)$$

in which φ represents the gravitational potential, x or y the horizontal axis, the ratio e/m is the ratio of a particle of mass m and charge e, and $-yA$, xA with $zA = 0$ the components of the magnetic field potential.

In order to maintain the neutrality of the ionized gas, an electric field with a potential ψ is set up to prevent the separation of positive and negative charges. In an atmosphere with uniform temperature $T_i = T_e = T$, the equation for each constituent is

$$\frac{dn_e}{dr} = \frac{n_e}{kT}\left(m_e \frac{d\Phi}{dr} + e \frac{d\psi}{dr}\right) \tag{44}$$

and

$$\frac{dn_i^+}{dr} = \frac{n_i^+}{kT}\left(m_i \frac{d\Phi}{dr} - Z_i e \frac{d\psi}{dr}\right) \tag{45}$$

Because the charge density is effectively zero, (42) is used and becomes, with (41), (44) and (45),

$$\Sigma Z_i n_i^+ \left(m_i \frac{d\Phi}{dr} - Z_i e \frac{d\psi}{dr}\right) = \Sigma Z_i n_i^+ \left(m_e \frac{d\Phi}{dr} + e \frac{d\psi}{dr}\right). \tag{46}$$

Hence, for a diffusion equilibrium at uniform temperature T,

$$e \frac{d\psi}{dr} = \frac{\Sigma Z_i n_i^+ (m_i - m_e)}{\Sigma Z_i (Z_i + 1) n_i^+} \frac{d\Phi}{dr} \tag{47}$$

represents the condition which keeps the gas neutral.

If μ_i the average mass for ions defined by

$$\mu_i = \frac{m_i + Z_i m_e}{Z_i + 1}, \tag{48}$$

condition (47) becomes

$$e \frac{d\psi}{dr} = \frac{d\Phi}{dz}\left[\frac{\Sigma Z_i (Z_i + 1) n_i^+ \mu_i}{\Sigma Z_i (Z_i + 1) n_i^+} - m_e\right]. \tag{49}$$

Since the first term of the bracket must represent the mean mass for ions, (49) has the following form

$$e \frac{d\psi}{dr} = \frac{d\Phi}{dz} (\mu_+ - m_e) \tag{50}$$

with

$$\mu_+ = \frac{\Sigma \, Z_i \, (Z_i + 1) \, n_i^+ \, \mu_i}{\Sigma \, Z_i \, (Z_i + 1) \, n_i^+}. \tag{51}$$

Thus, the fundamental equations (44) and (45) become with the neutrality condition (50)

$$\frac{dn_e}{n_e} = \frac{dr}{kT} \, \frac{d\Phi}{dr} \, \mu_+ \tag{52}$$

and

$$\frac{dn_i^+}{n_i^+} = \frac{dr}{kT} \, \frac{d\Phi}{dr} \, (m_i + Z_i m_e - Z_i \mu_+). \tag{53}$$

If the ionospheric gas consists of electrons together with _singly_ ionized atoms of various kinds, μ_+ given by (51) is written as

$$\mu_+ = \frac{\Sigma n_i^+ \, \mu_i}{\Sigma \, n_i^+} = \frac{\Sigma n_i^+ \, m_i}{2\Sigma n_i^+} + \frac{m_e}{2} \equiv \frac{m_+}{2} + \frac{m_e}{2}. \tag{54}$$

Hence the integrals of (52) and (53) are

$$n_e = n_{ea} \, \exp\left[\frac{1}{kT} \int_a^r \frac{d\Phi}{dr} \, \mu_+ dr \right] \tag{55}$$

and

$$n_i^+ = n_{ia}^+ \, \exp\left[\frac{1}{kT} \int_a^r \frac{d\Phi}{dr} \, (m_i + Z_i m_e - Z_i \mu_+) dr \right], \tag{56}$$

in which a = r represents the lowest level where equilibrium conditions are accepted, give the electron and ions distri-

butions. They become,with (54),

$$n_e = n_{ea} \exp\left[\frac{1}{2kT}\int_a^r \frac{d\Phi}{dr}(m_+ + m_e)\, dr\right] \qquad (57)$$

and

$$n_i^+ = n_{ia}^+ \exp\left[\frac{1}{kT}\int_a^r \frac{d\Phi}{dr}(m_i + m_e)dr - \frac{1}{2kT}\int_a^r \frac{d\Phi}{dr}(m_+ + m_e)dr\right].$$
$$(58)$$

The integral (58) becomes, with (57),

$$n_i^+ = \frac{n_{ea}}{n_e} n_{ia}^+ \exp\left[\frac{1}{kT}\int_a^r \frac{d\Phi}{dr}(m_i + m_e)\, dr\right], \qquad (59)$$

A similar formula for electrons is

$$n_e = \Sigma n_i^+ = \frac{n_{ea}}{n_e}\Sigma n_{ia}^+ \exp\left[\frac{1}{kT}\int_a^r \frac{d\Phi}{dr}(m_i + m_e)\, dr\right]. \qquad (60)$$

Hence,the general ionization equation is written as

$$\frac{n_e^2}{n_{ea}^2} = \frac{\Sigma n_{ia}^+ \exp\left[\frac{1}{kT}\int_a^r \frac{d\Phi}{dr}(m_i + m_e)\, dr\right]}{\Sigma n_{ia}^+} \qquad (61)$$

Equations (59) and (61) represent the concentrations of ions and electrons in a steady state at uniform temperature when the ionized gas involves electrons with singly ionized atoms. Such a steady state corresponds to a diffusion equilibrium with boundary conditions fixed at $r = a$ by concentrations n_{ea} and n_{ia}^+ .

If the potential function Φ is such as

$$\frac{d\Phi}{dr} = -g = g_a(a/r)^2 \tag{62}$$

the fundamental formulas (59) and (61) are written as

$$n_i^+/n_{ia}^+ = (n_{ea}/n_e) \exp -\left[\frac{g_a(m_i + m_e)}{kT}\left(\frac{a}{a+z}\right)z\right] \tag{63}$$

and

$$\frac{n_e^2}{n_{ea}^2} = \frac{\Sigma n_{ia}^+ \exp -\left[\frac{g_a(m_i + m_e)}{kT}\left(\frac{a}{a+z}\right)z\right]}{\Sigma n_{ia}^+} \tag{64}$$

where $z = r - a$.

Since m_e is negligible compared with m_i (64) leads to

$$\frac{n_e}{n_{ea}} = \left[\frac{\Sigma n_{ia}^+ (n_i/n_{ia})}{\Sigma n_{ia}^+}\right]^{1/2} \tag{65}$$

where n_i/n_{ia} represents the vertical distribution in diffusive equilibrium of neutral atoms of mass m_i.

If there is only one kind of ion of charge $Z = 1$, (65) leads to

$$\frac{n_1^+}{n_{1a}} = \frac{n_e}{n_{ea}} = \left[\frac{n_1}{n_{1a}}\right]^{1/2} = \exp\left[-\left(\frac{a}{a+z}\right)\frac{m\,g_a}{2kT}z\right]; \tag{66}$$

that is, the scale height H_+ of the vertical distribution of ion and electron concentrations is twice the scale height of the neutral constituent.

The formula (63) or

$$\frac{n_i^+}{n_{ia}^+} = \frac{n_{ea}}{n_e} \frac{n_i}{n_{ia}} \tag{67}$$

implies that $Z_i = 1$, and cannot be applied to a highly charged ion. Using (56) for He^+ and O^{++}, we see that mixtures consisting of electrons with light ions (He^+) or with heavy ions which are lightly charged have different behaviors. Namely,

$$\frac{n^+(He)}{n_a^+(He)} = \frac{n_{ea}}{n_e} \frac{n(He)}{n_a(He)} \tag{68}$$

$$\frac{n^+(O)}{n_a^+(O)} = \frac{n_{ea}}{n_e} \frac{n(O)}{n_a(O)} \tag{69}$$

$$\frac{n^{++}(O)}{n_a^{++}(O)} = \left[\frac{n_{ea}}{n_e}\right]^2 \frac{n(O)}{n_a(O)} \tag{70}$$

$$\frac{n^{+++}(O)}{n_a^{+++}(O)} = \left[\frac{n_{ea}}{n_e}\right]^3 \frac{n(O)}{n_a(O)} \tag{71}$$

In a layer where He^+ is the principal ion, very nearly,

$$\frac{n^+(He)}{n_a^+(He)} = \left[\frac{n(O)}{n_a(O)}\right]^{1/4} \tag{72}$$

In the same conditions, for O^{++},

$$\frac{n^{++}(O)}{n_a^{++}(O)} = \left[\frac{n(O)}{n_a(O)}\right]^{1/3} \tag{73}$$

and for O^{+++}

$$\frac{n^{+++}(0)}{n_a^{+++}(0)} = \left[\frac{n(0)}{n_a(0)}\right]^{1/4} \tag{74}$$

It is also of interest to estimate the effect of temperature in an ionospheric layer composed mainly of oxygen ions O^+ and helium ions He^+. Since $m(O) = 4m(He)$, (68) and (69) give the following ratio

$$\frac{n^+(He)}{n^+(0)} = \frac{n_a^+(He)}{n_a^+(0)} \left[\frac{n_a(0)}{n(0)}\right]^{3/4}, \tag{75}$$

showing that it must increase at each level when the temperature decreases.

Figure 6 shows the variation of such a ratio for various temperatures between $2000^{\circ}K$ and $1000^{\circ}K$. The effect of temperature is very important. Furthermore, if the two scale heights $H(He)$ and $H(He^+)$ are compared, their ratio is written, (53) and (54),

$$\frac{H(He)}{H(He^+)} = 1 - \frac{m_+}{2m(He)} \tag{76}$$

$$H(He^+) = 2H(He) \frac{n^+(0) + n^+(He)}{n^+(He) - 2\,n^+(0)} \tag{77}$$

which shows that helium ions are distributed according to a negative scale height where

$$n^+(He) < 2n^+(0) . \tag{78}$$
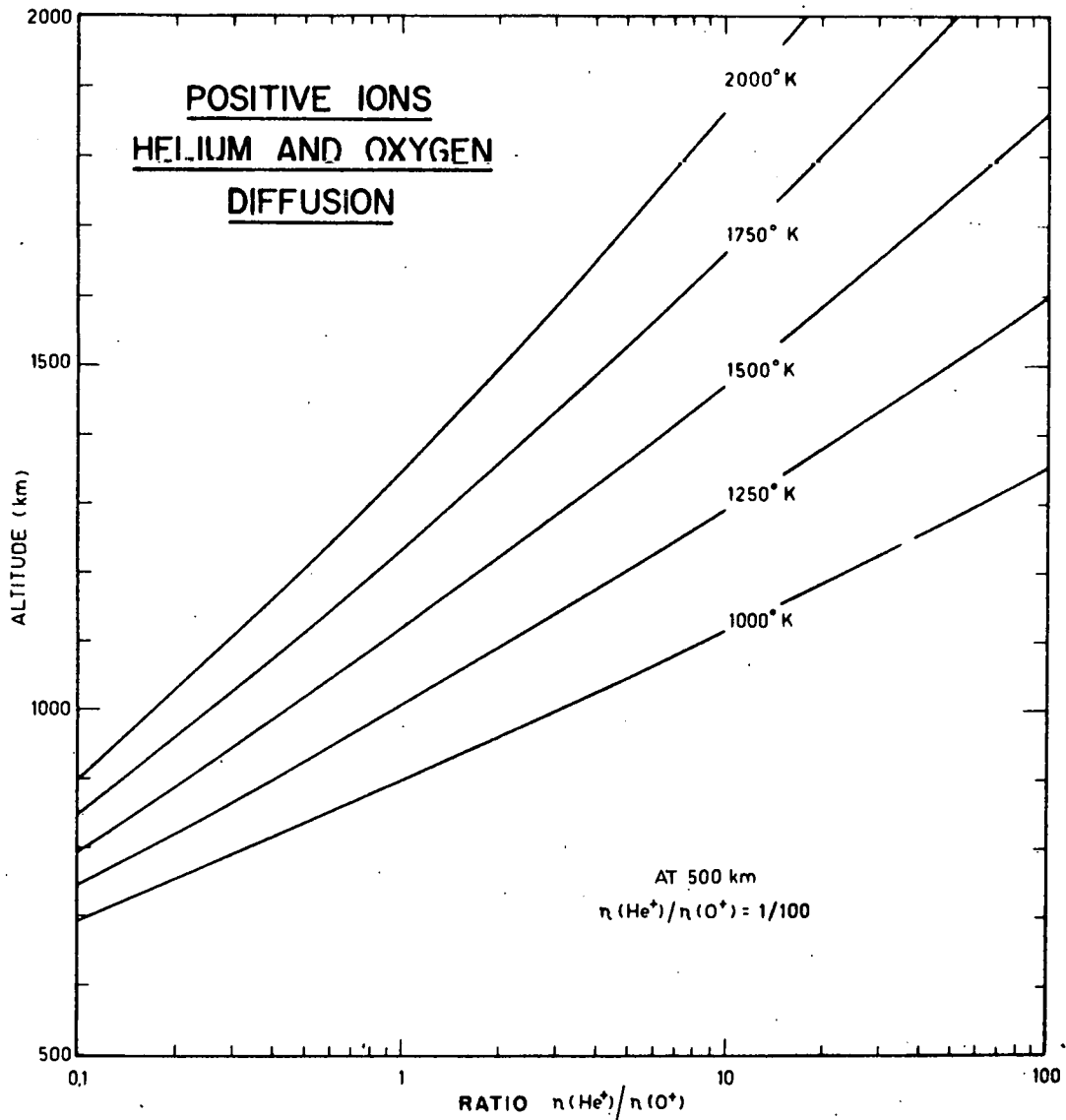
VARIATION OF THE RATIO OF He⁺ AND O⁺ CONCENTRATIONS
WITH TEMPERATURES FOR BOUNDARY CONDITIONS AT 500 KM
$n(He^+)/n(O^+) = 10^{-2}$

FIGURE 6

If $n^+(He)$ is negligible compared with $n^+(O)$, the scale height of He^+ is

$$H(He^+) = - H(He); \tag{79}$$

that is the scale height of the minor ion He^+ has an absolute value corresponding to that of the neutral atom. There is still an increase of the concentration of the light ion with altitude where $n^+(He) = n^+(O)$ since $H(He^+)$ is given by

$$H(He^+) = - 4 H(He) . \tag{80}$$

The condition

$$n^+(He) < 2n^+(O), \tag{81}$$

leads to a positive scale height of He^+ and its concentration will begin to decrease with height.

To summarize: in an ionospheric region where (63) and (64) represent atmospheric conditions, the general equation for each species of ions is

$$\frac{n^+(X) \, n_e}{n(X)} = \frac{n_a^+(X) \, n_{ea}}{n_a(X)} . \tag{82}$$

Hence, for two ions Y^+ and Z^+ with X^+

$$\frac{n^+(X) \, n(Y)}{n^+(Y) \, n(X)} = \frac{n_a^+(X) \, n_a(Y)}{n_a^+(Y) \, n_a(X)} \tag{83}$$

and

$$\frac{n^+(X) \, n(Z)}{n^+(Z) \, n(X)} = \frac{n_a^+(X) \, n_a(Z)}{n_a^+(Z) \, n_a(X)} . \tag{84}$$

These three equations clearly show how the vertical distribution of ionized constituents depends on the conditions at the lower boundary. Such conditions are related to the photoionization and recombination processes or to reactions controlling the ratio of ions at the reference level.

A direct application of the last condition is the charge transfer process between hydrogen and oxygen,

$$H + O^+ \rightleftharpoons H^+ + O . \tag{85}$$

If such a process controls the ionization at the reference level $r = a$, one can write the following condition

$$\frac{n_a^+(O) \, n_a(H)}{n_a^+(H) \, n_a(O)} = \frac{8}{9} \tag{86}$$

where the ratio 8/9 is the ratio of products of statistical weights. According to (83) such a condition is satisfied everywhere when (63) is accepted. As has been shown before, the ratio $n_a(O)/n_a(H)$ of the neutral constituents is very sensitive to temperature. There is a variation of a factor of about 100 of such a ratio when the temperature varies between $1000^{\circ}K$ and $2000^{\circ}K$. Hence the variation of the ratio $n_a^+(O)/n_a^+(H)$ of ions is almost identical to that of the ratio of neutral atoms according to (86); an immediate result is that H^+ becomes more important at low than at high temperatures. Furthermore, the life-time of neutral hydrogen at 500 km, being of the order of a few hours, a diurnal variation

of $n(H)$ must occur and such a variation leads to a variation of $n^+(H)$ according to (83). As an example, Fig. 7 shows how vertical distributions of O^+, He^+ and H^+ vary when the temperatures are between $1000^\circ K$ and $2000^\circ K$. With an arbitrary concentration of atomic oxygen ions of the order of 10^5 cm^{-3} in the neighborhood of 500 km, for which $n^+(He)$ is about 500 cm^{-3}, the variation of $n_a^+(H)$ is about a factor of 100. At highest temperatures, a layer of O^+ ions is maintained up to a belt of helium ions. At lowest temperatures, there is practically no transition layer between the belt of oxygen ions and belt of hydrogen ions.

Finally, it should be pointed out that, in any planetary atmosphere, one reaches steady state conditions only in special circumstances and that one cannot always assume equilibrium conditions to a sufficiently good approximation. It can easily be seen, then, that dynamical processes must be introduced to explain real physical conditions.
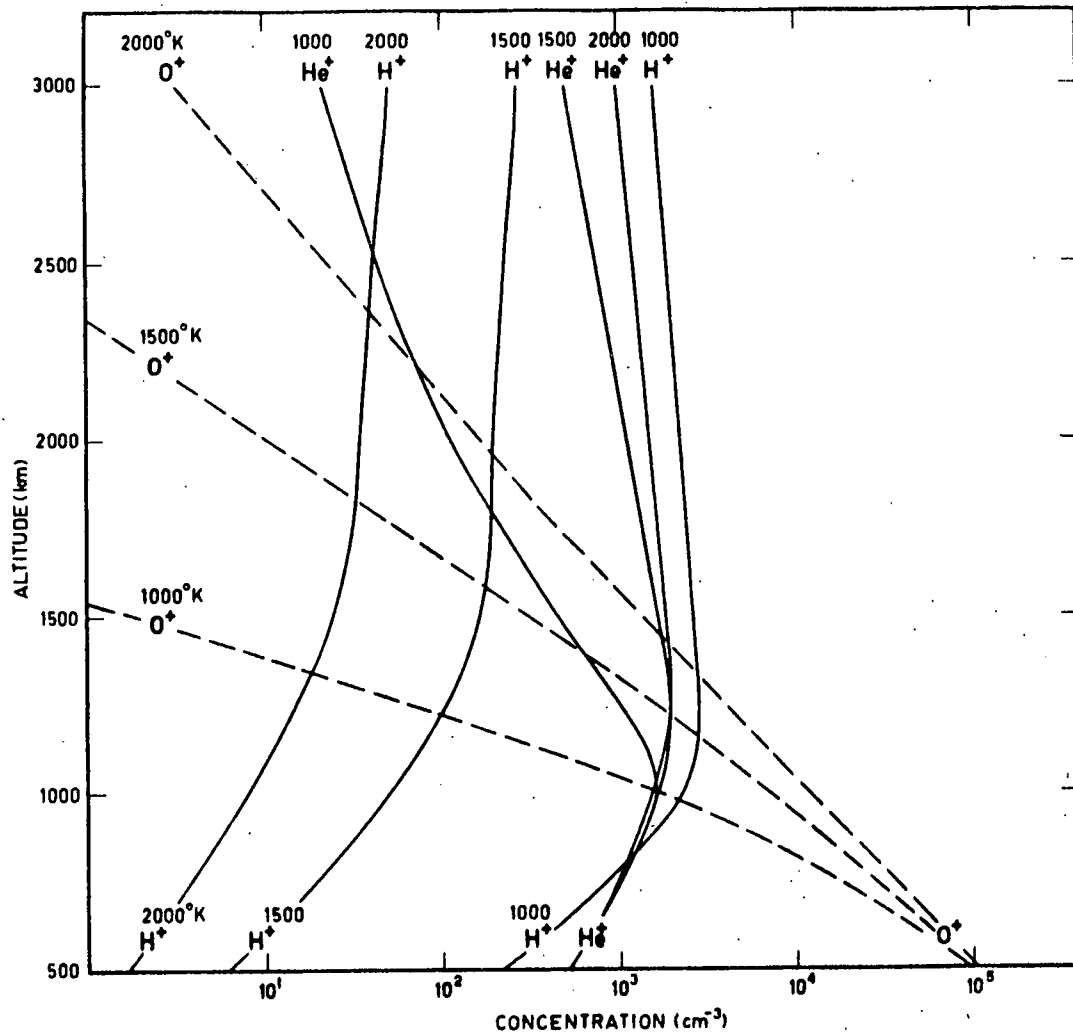
VERTICAL DISTRIBUTION OF O^+, He^+ AND H^+ IONS IN DIFFUSION EQUILIBRIUM FOR BOUNDARY CONDITIONS DEFINED AT 500 Km. THE TEMPERATURE EFFECT IS VERY IMPORTANT FOR HYDROGEN IONS AND AFFECTS THE DISTRIBUTION OF OTHER IONS.

FIGURE 7

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