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Helium and hydrogen distributions in the upper atmosphere

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FOREWORD

This text has been presented at the International School of Atmospheric Physics which was held in Erice (Trapani), Italy, 15 - 29 June 1970. It will be published in the proceedings of the school.

AVANT-PROPOS

Ce texte a été présenté lors de l'Ecole Internationale de Physique Atmosphérique qui a eu lieu à Erice (Trapani) Italie du 15 au 29 juin 1970. Il sera publié dans le volume reprenant les divers exposés.

VOORWOORD

Deze tekst werd voorgedragen ter gelegenheid van de Internationale School voor Atmosferische Fysica die plaats had te Erice (Trapani) Italië van 15 tot 29 juni 1970, en zal gepubliceerd worden in een uitgave die al de gegeven uiteenzettingen bevat.

VORWORT

Dieser Text wurde während der Internationalen Sommerschule über atmosphärische Physik in Erice (Trapani), Italien vom 15. bis zum 29. Juni 1970, vorgestellt. Er wird in einem Buch mit den anderen Darlegungen herausgegeben werden.

HELIUM AND HYDROGEN DISTRIBUTIONS IN THE UPPER ATMOSPHERE

by

G. KOCKARTS.

Abstract

The physical processes involved in the computation of helium and hydrogen distributions are described for steady state conditions. A solution of the general transport equation is given to indicate the effects of molecular and eddy diffusion. The concept of maximum transport velocity is introduced in relation with the initial distribution prevailing around 100 km altitude. The effect of eddy diffusion is particularly important for the helium distribution. Variations of a factor of 4 in the helium concentration at high altitude can be explained by a change of factor of 10 for the eddy diffusion coefficient in the 100 km region.

Résumé

Les processus physiques impliqués dans le calcul des distributions de l'hélium et de l'hydrogène sont décrits pour des conditions stationnaires. La solution générale de l'équation de transport indique les effets de la diffusion moléculaire et de la diffusion turbulente. La notion de vitesse maximale de transport est introduite en relation avec la distribution initiale existant au voisinage de 100 km d'altitude. L'effet de la diffusion turbulente est très important pour la distribution de l'hélium. Un changement d'un facteur 10 dans la valeur de coefficient de diffusion turbulente aux environs de 100 km permet d'expliquer des variations d'un facteur 4 de la concentration de l'hélium à haute altitude.

Samenvatting

Vooreerst worden de fysische processen beschreven die de grondslag vormen voor de berekening van de verdeling van helium en waterstof in een stationaire toestand. Vervolgens wordt een oplossing gegeven voor de algemene transportvergelijking waarbij rekening gehouden werd met moleculaire en turbulente diffusie. Het begrip van maximale transportsnelheid wordt ingevoerd in verband met de oorspronkelijke verdeling op 100 km hoogte. De invloed van turbulente diffusie is zeer belangrijk voor de heliumdistributie. Viervoudige wijzigingen van de heliumconcentratie op grote hoogten kunnen verklaard worden door aan te nemen dat de turbulentediffusiecoëfficiënt met een factor 10 gewijzigd wordt op een hoogte van een honderdtal km.

Zusammenfassung

Die physikalische Vorgänge, die in der Rechnung der Helium und Wasserstof Verteilungen einbegriffen sind, werden für stationären Zuständen beschrieben. Eine generale Lösung der Transportgleichung bestimmt die Wirkung der Molekular - und Eddy - Diffusion. Die maximale Transportgeschwindigkeit wird für die Anfangsverteilung in der Umgebung von 100 km beschrieben. Die Eddydiffusion ist sehr wichtig für der Heliumverteilung. Wenn der Eddydiffusionskoeffizient bei einem Faktor 10 in der Umgebung von 100 km verändert wird, so kann die Heliumkonzentration bei einem Faktor 4 ändern.

1.- INTRODUCTION

The study of the upper atmosphere above 100 km has developed greatly, since the first rocket soundings in 1945. It has been possible to show that the atmospheric density decreases less between 100 km and 200 km than between ground level and 100 km. The diffusion effect in the Earth's gravity field was theoretically derived by Nicolet and Mange (1954) and has been confirmed by the observation of molecular oxygen up to 180 km altitude (Byram, Chubb and Friedman, 1957). Observations of the atomic hydrogen Lyman-α nightglow (Kupperian, Byram, Chubb and Friedman, 1958) indicated that atomic hydrogen can play a significant role above 100 km altitude. These results imply that there must be a change in the composition of the upper atmosphere (Mange, 1955 ; Nicolet, 1960 ; Mange, 1961).

Since 1957, considerable progress has been accomplished on the density distribution deduced from satellite drag data. Even with round figures, it is possible to show that the decrease in density at high altitude is completely different from the decrease observed between ground level and 100 km height. Within the first hundred kilometers the density decreases roughly from 10^{-3} gm cm^{-3} to 10^{-9} gm cm^{-3}, whereas between 150 km and 400 km, between 400 km and 700 km and between 700 km and 1800 km the decrease is successively of the order of a factor 100.

A general understanding of this situation can be obtained from the hydrostatic equation and from the perfect gas law.

These equations lead to the general relation :

$$\frac{d\rho}{\rho} + \frac{dT}{T} - \frac{dm}{m} = - \frac{dz}{kT/mg} \qquad (1)$$

where ρ, T and m respectively represent the density, the temperature and the mean molecular mass at a height z where g is the gravity acceleration. Equation (1) shows that the slow decrease of the density is due to an increase of the atmospheric scale height H given by

$$H = kT/mg \tag{2}$$

where k is Boltzmann's constant.

An increase of H with height can be due to two factors : an increase of the temperature T and/or a decrease of the mean molecular mass m. The temperature increase is directly related to solar ultraviolet heating and to heat transport. A decrease in the mean molecular mass can arise from a diffusive separation of the atmospheric constituents. At ground level a mean molecular mass of 29 corresponds to an atmospheric composition of 21% O_2, 78% N_2 and 1% Ar. If diffusion prevails above a certain height, the mean molecular mass can a priori reach a value of 1 amu (atomic hydrogen) after passing through the values corresponding to atomic oxygen (16 amu) and to helium (4 amu). The helium mixing ratio is 5.24×10^{-6} (Gluckauf 1946) at ground level out can reach values of the order of 1 in the upper atmosphere.

The problem is even more complex for hydrogen which does not exist in its atomic form at ground level. Below 100 km atomic hydrogen is produced through photodissociation of water vapor and methane as well through oxidation reactions of excited $O(^1D)$ atoms with CH_4, H_2 and H_2O (Nicolet, 1970). These minor constituents are subject to molecular and eddy diffusion below 100 km and an atomic hydrogen concentration of 3×10^7 cm^{-3} corresponds to a relative abundance of the order of 2×10^{-6} at 100 km altitude. At greater heights, this relative abundance also increases and diffusive transport creates successively an atomic oxygen belt, a helium belt and an atomic hydrogen belt. In this paper, the physical processes responsible for the increase of the He and H relative abundances will be described under steady state conditions.

2.- DIFFUSION EQUATION

When an atmospheric component is transported by diffusion in the atmosphere, it is possible to express the flow $F = n_1 w$ of particles 1 in the form

$$F = F_D + F_E \tag{3}$$

where F_D represents the flux due to molecular diffusion and F_E is the flux due to eddy diffusion.

The molecular diffusion flow tends to distribute the different species according to their own mass, whereas eddy diffusion tends to maintain a mixing distribution in which all atmospheric constituents are distributed according to the mean molecular mass. Chemical reactions, photodissociation and ionization processes can compete with diffusive transport. If these processes are fast enough, a particular atmospheric constituent will have a vertical distribution independent of molecular or eddy diffusion effects. For example, in the daytime mesosphere ozone is always in photoequilibrium, since photodissociation is rapid and only a time of the order of 100 sec is required to reach the photoequilibrium conditions. This means that the times necessary to reach molecular diffusion equilibrium or mixing conditions are too long compared to the photodissociation time for diffusion to be important. This is not the case for helium and hydrogen in the thermosphere however.

For a two component gas, the relative molecular diffusion velocity $(w_1 - w_2)$ can be written (Chapman and Cowling, 1939)

$$w_1 - w_2 = - D_{12} \left[\frac{n^2}{n_1 n_2} \frac{d(n_1/n)}{dz} + \frac{m_2 - m_1}{m} \frac{1}{p} \frac{dp}{dz} + \frac{\alpha}{T} \frac{dT}{dz} \right] \tag{4}$$

In this equation, the subscripts 1 and 2 refer to the component 1 and 2, respectively. D_{12} is the molecular diffusion coefficient, p is the total pressure, α is the thermal diffusion factor and $n = n_1 + n_2$ is the total concentration. The other symbols have the same meaning as in equations (1) and (2). From equation (4) it is seen that molecular diffusion arises from three different effects ; namely, concentration, pressure and temperature gradients.

If it is assumed that there is no net flow through an elementary surface at height z, i.e. if

$$n_1 w_1 + n_2 w_2 = 0 \tag{5}$$

then equation (4) can be transformed into

$$w_1 = - D_{12} \left[\frac{1}{n_1} \frac{dn_1}{dz} - \frac{1}{n} \frac{dn}{dz} + \frac{(m_2 - m_1)}{m\,n} \frac{n_2}{p} \frac{dp}{dz} + \alpha \frac{n_2}{n} \frac{1}{T} \frac{dT}{dz} \right] \tag{6}$$

Using the hydrostatic equation and the perfect gas law, equation (6) can be written

$$w_1 = - D_{12} \left[\frac{1}{n_1} \frac{dn_1}{dz} + \frac{1}{H_1} + (1 + \alpha \frac{n_2}{n}) \frac{1}{T} \frac{dT}{dz} \right] \tag{7}$$

where $H_1 = kT/m_1 g$ is the pressure scale height of the constituent 1.

As helium and hydrogen are minor constituents in the lower thermosphere, their concentration and density are negligible compared to the total concentration n and to the total density ρ. In this case, equation (7) becomes

$$w_1 = - D_{12} \left[\frac{1}{n_1} \frac{dn_1}{dz} + \frac{1}{H_1} + (1 + \alpha) \frac{1}{T} \frac{dT}{dz} \right] \tag{8}$$

A binary diffusion equation is used here since the minor constituent is assumed to diffuse through another component which is actually the sum of all the major atmospheric constituents (O, O_2 and N_2). The molecular diffusion coefficient is therefore given by

$$D_{12} = D_{1M} = \frac{3\pi}{32Q} \left(1 + \frac{m_1}{m}\right)^{1/2} \left(\frac{8kT}{\pi m_1}\right)^{1/2} \frac{1}{n} \qquad (9a)$$

where Q is a momentum transfer cross section, m is the mean molecular mass variable with height and n is the total concentration. For a momentum transfer cross section of 2.83×10^{-15} cm^2, the molecular diffusion coefficient (9a) is given by

$$D_{1M} = 1.96 \times 10^6 \left(\frac{1}{m_1} + \frac{1}{m}\right)^{1/2} \frac{T^{1/2}}{n} \ (\text{cm}^2 \ \text{sec}^{-1}) \qquad (9b)$$

if the masses m_1 and m are expressed in gm, n in cm^{-3} and T in °K. Fig. 1 shows the vertical distribution of the molecular diffusion coefficient for helium between 85 km and 120 km. The total concentration is also represented on the same figure which indicates the large increase of $D_{He,M}$ proportionally to the decrease of $n = \Sigma n_i$.

The effect of thermal diffusion on the helium vertical distribution has been discussed by Kockarts (1963) and the thermal diffusion factor α can be taken equal to -0.38 for helium and hydrogen. The thermal diffusion factor actually depends on the relative abundance of the diffusing component (Chapman and Cowling, 1952), but in the region where thermal diffusion is found to be important, helium and hydrogen are always minor constituents. Under these conditions, the interaction potential between the minor and the major component is not as important as when both constituents are of the same order of magnitude. The value -0.38 is then obtained for a rigid sphere interaction potential between the minor constituent (He or H) and molecular nitrogen which is the major component in the region where important temperature gradients exist.

Following Lettau (1951), the vertical eddy diffusion velocity w_E can be written

$$w_E = -K \left. \frac{d(n_1/n)}{dz} \middle/ (n_1/n) \right. \qquad (10)$$

where K is the eddy diffusion coefficient.

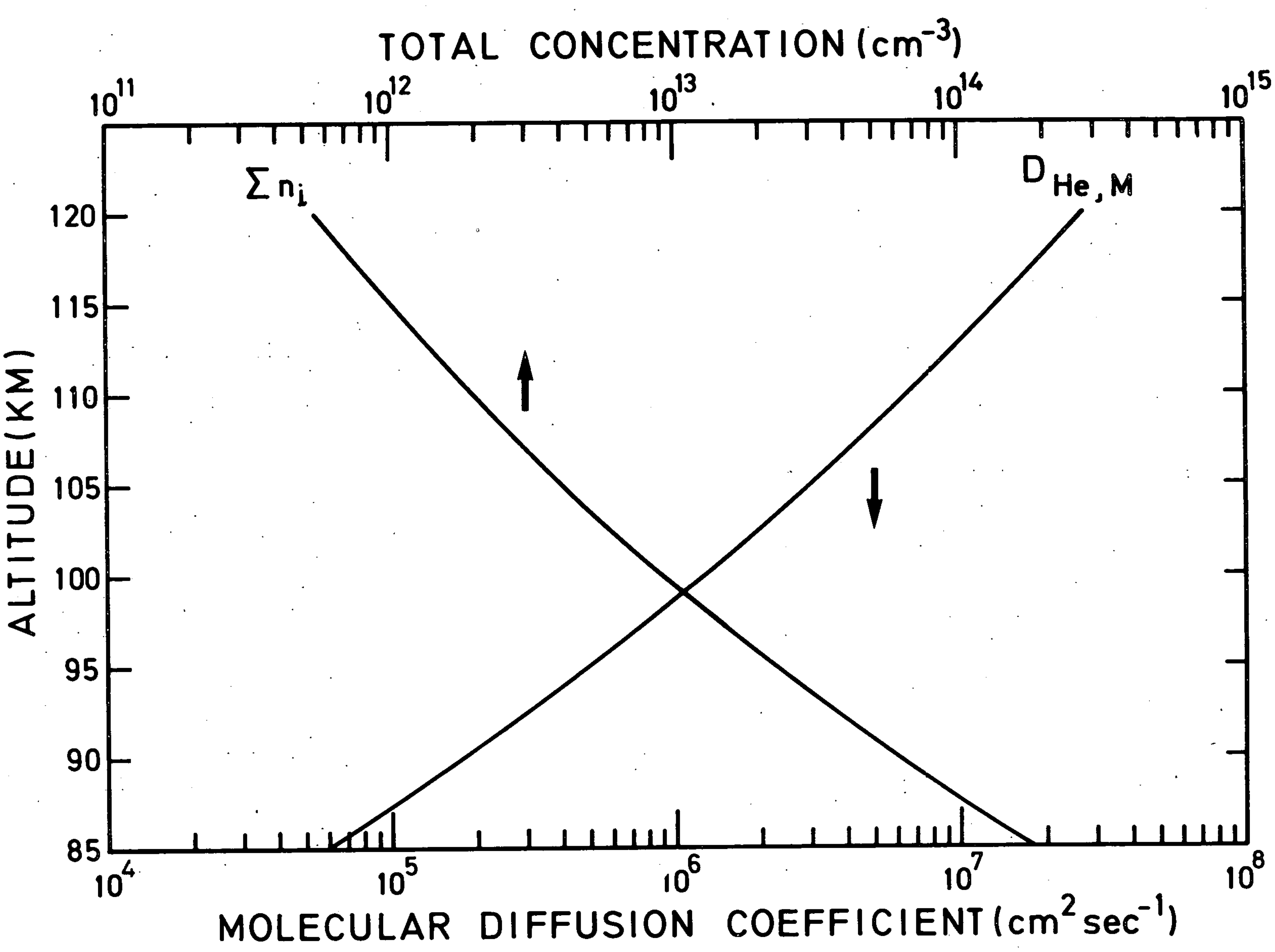

Fig. 1.- Vertical distribution of the molecular diffusion coefficient for helium and of the total concentration.

Using the general relation

$$\frac{dn}{n} = - \frac{dz}{H} - \frac{dT}{T} \tag{11a}$$

which arises from the hydrostatic equation

$$\frac{dp}{dz} = - \rho g \tag{11b}$$

equation (10) becomes

$$w_E = - K \left[\frac{1}{n_1} \frac{dn_1}{dz} + \frac{1}{H} + \frac{1}{T} \frac{dT}{dz} \right] \tag{12}$$

If we introduce a parameter $\Lambda \equiv K/D$, equation (3) can be transformed into the following expression, by using (8) and (12) :

$$\frac{dn_1}{dz} + \left[\frac{1}{1+\Lambda} \left(\frac{1}{H_1} + \frac{\Lambda}{H} \right) + \left(1 + \frac{\alpha}{1+\Lambda} \right) \frac{1}{T} \frac{dT}{dz} \right] n_1 = - \frac{F}{D_{12}(1+\Lambda)} \tag{13}$$

This expression is similar to the form proposed by Colegrove et al. (1966) for a multicomponent system. The only difference results from the expression for the diffusion coefficient D_{12}. Colegrove et al. (1966) use an expression such that $D_{12}^{-1} = \sum_j D_{1j}^{-1}$, where D_{1j} are the binary diffusion coefficients. It is interesting to note, however, that for the atmospheric models used in the present work expression (9b) and the relation $D_{12}^{-1} = \sum_j D_{1j}^{-1}$ yield the same numerical results if a common momentum transfer cross section is used for all the binary diffusion coefficients. When $\Lambda = 0$, i.e. when eddy diffusion is neglected, equation (13) reduces to the relation used by Kockarts and Nicolet (1962, 1963) for the computation of atomic hydrogen distributions. It will be shown later that the introduction of an eddy diffusion coefficient is more important for helium than for atomic hydrogen.

A general solution of equation (13) can now be written in the following form

$$n_1(z) = n_{eq}\left[1 - \int_{z_1}^{z} \frac{F}{D_{12}(1+\Lambda)n_{eq}}\, dz'\right] \qquad (14)$$

where z_1 is the altitude at which the concentration n_1 is given and n_{eq} is the solution of equation (13) when the flow F is zero. The equilibrium distribution n_{eq} will be discussed later. A solution of the same type as (14) has been presented by Donahue (1969) in an analysis of deuterium in the upper atmospheres of Venus and Earth.

3.- TRANSPORT VELOCITIES

From equation (14), it is immediately seen that any departure from an equilibrium solution is related to the importance of the transport flow F compared to $D_{12}(1+\Lambda)$. When a siginficant flow is present, the corresponding concentration distribution is always less than the equilibrium distribution. It is therefore necessary to investigate the transport velocities associated with equations (13) or (14).

We consider first the physical conditions prevailing in the lower boundary region, where there is a tendency to diffusive separation. It is always possible to express the concentration gradient of a minor constituent as a function of the total concentration gradient in a form such as

$$\frac{1}{n_1}\frac{dn_1}{dz} = X(z)\,\frac{1}{n}\frac{dn}{dz} \qquad (15)$$

where X(z) is the vertical distribution factor which is height dependent. Using relation (11), equation (15) can be written

$$\frac{1}{n_1}\frac{dn_1}{dz} = -X(z)\left[\frac{1}{H} + \frac{1}{T}\frac{dT}{dz}\right] \qquad (16)$$

With the preceeding equation, the total transport velocity w can be obtained from equation (13)

$$w = \frac{D_{12}}{H}\left[X - \frac{H}{H_1} - \frac{H}{T}\frac{dT}{dz}(1 + \alpha - X)\right] + \frac{K}{H}\left[X - 1\right] \quad (17)$$

where the contributions due to molecular diffusion and to eddy diffusion are respectively represented by the terms proportional to D_{12} and to K.

When the minor constituent n_1 is characterized by a mixing distribution, the vertical distribution factor is equal to one in the region where mixing prevails. Under this condition, the transport velocity arises from molecular diffusion alone and equation (17) reduces to

$$w = \frac{D_{12}}{H}\left[1 - \frac{H}{H_1} - \frac{\alpha H}{T}\frac{dT}{dz}\right] \quad (18)$$

Actually equation (18) represents the maximum molecular diffusion transport velocity for a constituent which follows the distribution of the major components. When the mass m_1 of the minor constituent is less than the mean molecular mass m, equation (18) shows that w is positive, i.e. upwards. This is the case for helium and hydrogen. In the case of Ar however, $w < 0$, i.e. downwards, since $m_1 > m$. Table I gives the maximum transport velocities between 100 km and 120 km for hydrogen, helium and argon. It can be seen that the molecular diffusion velocities are always of the order of a few cm $\sec^{-1}$ around 100 km altitude. A knowledge of the concentration at the lower boundary gives thus immediately the maximum flow which can be transported by molecular diffusion into the upper atmosphere.

TABLE I.- Maximum molecular diffusion transport velocities (cm sec^{-1})

z(km)	H(km)	T(°K)	w_H	w_{He}	w_{Ar}
100	6.41	209.2	3.76	1.78	- 0.36
105	7.24	230.9	8.35	3.94	- 0.84
110	8.42	261.9	15.6	7.35	- 1.72
115	9.63	293.0	28.1	13.3	- 3.31
120	10.87	324.0	48.8	23.0	- 6.11

Since the molecular diffusion velocity is proportional to D_{12}, it increases with height. Above a certain level however, the diffusion equation can no longer been applied since collisions become less frequent. This is the beginning of the exosphere where collisions are neglected. Actually the term F/D_{12} in equation (13) becomes less important with increasing height and the diffusive flow is controlled by escape from the Earth's gravitational field. If r represents the geocentric height of the critical level, i.e. the beginning of the exosphere, it is possible to show that, at this level, the effusion velocity is given by (see for example Nicolet, 1957)

$$v = (g/2\pi)^{1/2}\, r\, H_1^{-1/2}\, (1 + H_1/r)\, e^{-r/H_1} \qquad (19)$$

Fig. 2 shows the effusion velocity of atomic hydrogen as a function of the thermopause temperature. The two curves correspond to critical level fixed at 500 km and 1000 km. The critical level is actually temperature dependent, but this is not important for the present discussion. It is clear from Fig. 2 that v varies more strongly with temperature then with height. Fig. 3 shows the same situation for deuterium and here the importance of the mass involved in H_1 is evident. For helium, the effusion velocity is always less then 1 cm sec^{-1} even for temperatures as high as 2000°K. Fahr (1970) discussed some departures from expression (19), due to ballistic transport phenomena. Other processes can also lead to slight

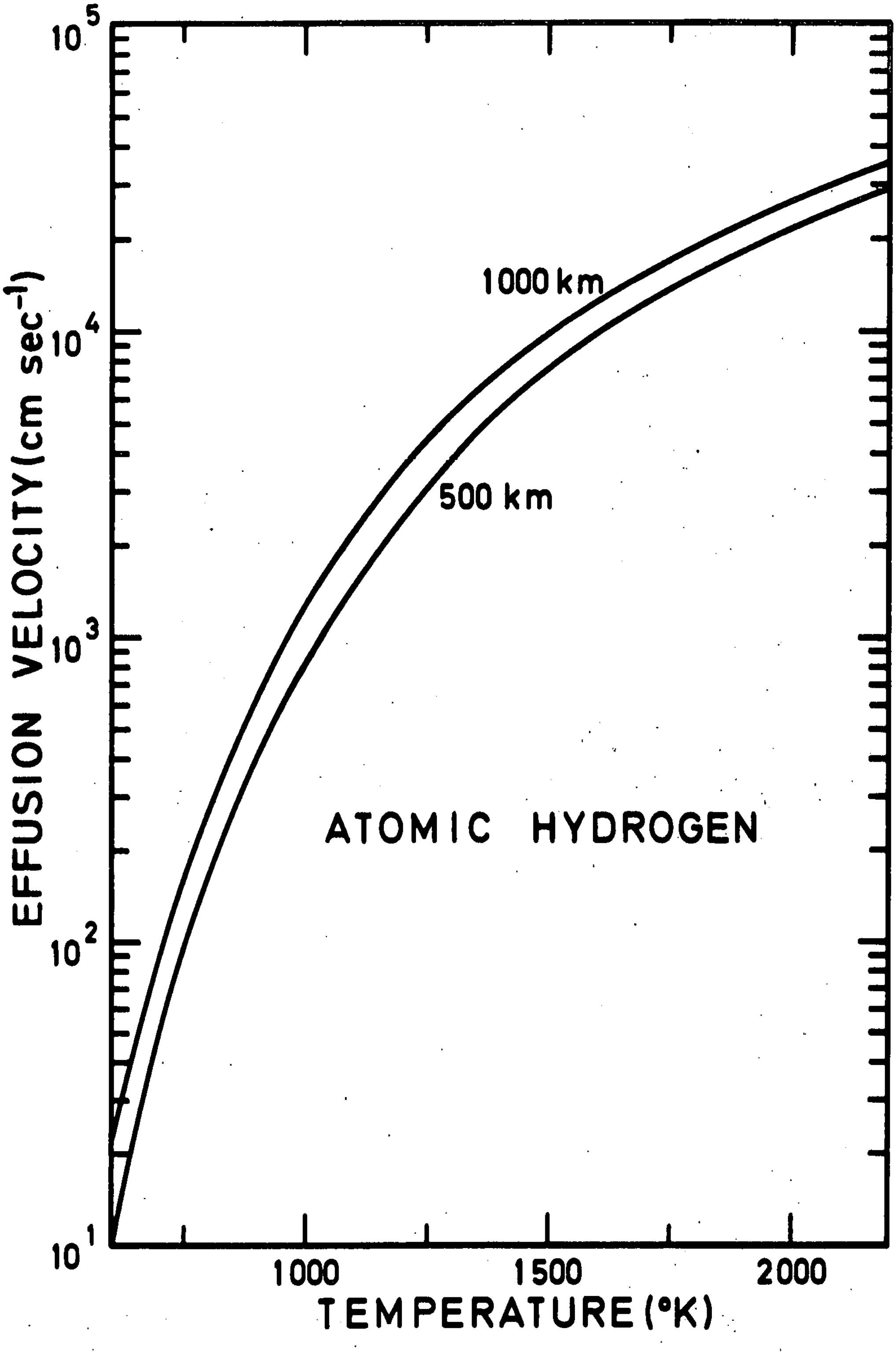

Fig. 2.- Effusion velocity of atomic hydrogen as a function of the thermopause temperature for two critical levels fixed at 500 km and 1000 km respectively.

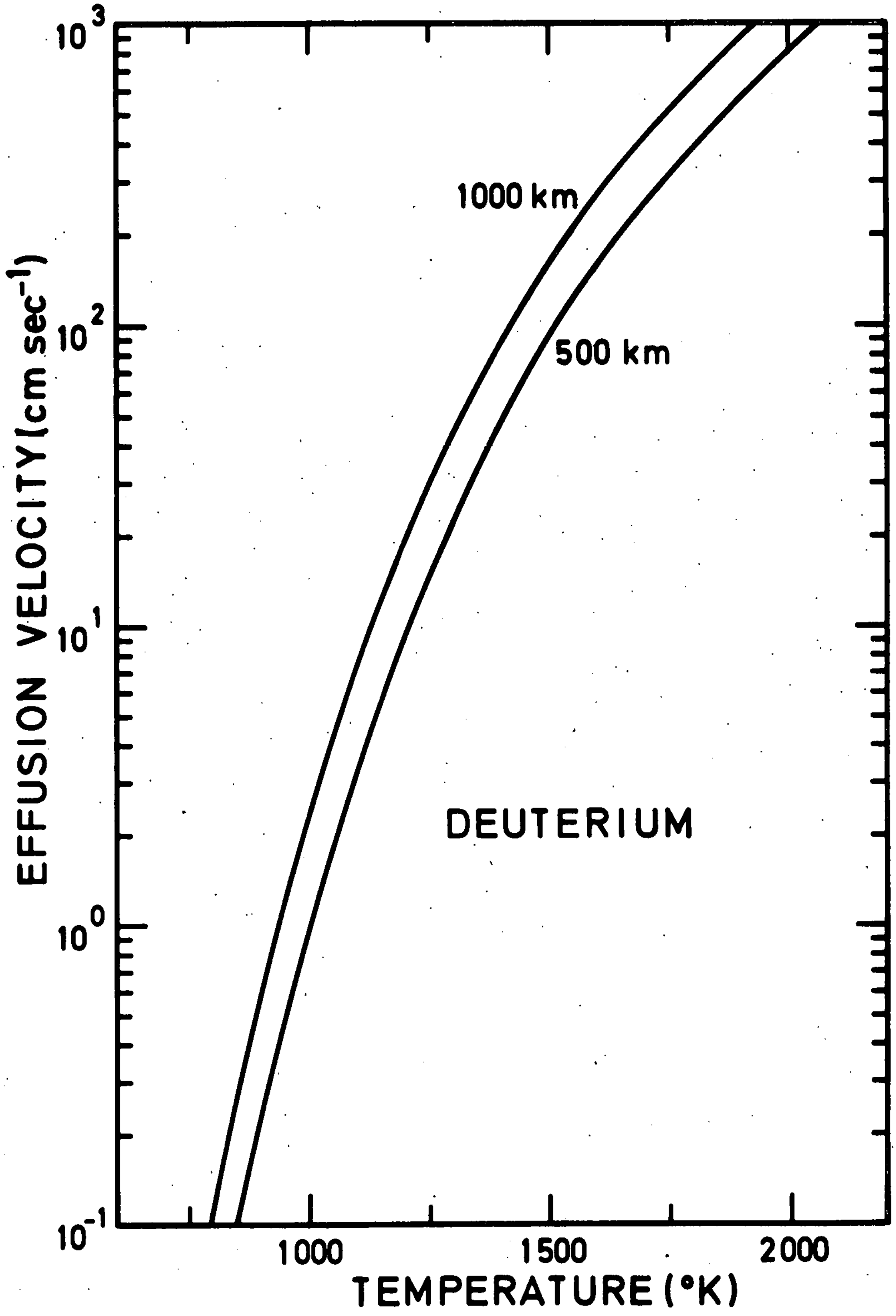

Fig. 3.- Effusion velocity of deuterium as a function of the thermopause temperature for two critical levels fixed at 500 km and 1000 km respectively.

differences with (19) and the interested reader is referred to the paper by Brinkmann (1970). It should only be mentioned that corrections of the order of 10 to 20% to equation (19) do not modify the physical analysis presented here, since at the present time the concentration of atomic hydrogen is not known with an accuracy better than 20%.

Since the light particles are escaping from the Earth's gravity field, it is possible to introduce the flow condition

$$F = n_1(r)\ v(r) \tag{20}$$

at the critical level.
Without loss or production in the upper atmosphere where equation (13) is applied, the flux at two levels r_1 and r_2 must satisfy the condition

$$F_1\ r^2_1 = F_2\ r_2^2 \tag{21}$$

which arises from the equation of continuity.
With equations (19), (20) and (21), it is then possible to compute the concentration distribution using (14).

4.- VERTICAL DISTRIBUTIONS AND EDDY DIFFUSION

For a hydrogen concentration of the order of 4×10^7 cm^{-3} at 100 km, the maximum molecular diffusion flow is of the order of 1.5×10^8 cm^{-2} sec^{-1}, according to Table I. A hydrogen concentration of such an order of magnitude is required to explain the Lyman-α scattering observations (Donahue, 1966 ; Tinsley, 1969 ; Meier, 1970). Thus, the real upwards flow through the lower thermosphere is practically identical to the maximum possible flow permitted by molecular diffusion. Solutions of equations (13) or (14) are presented in Fig. 4 for several thermopause temperatures. These vertical distributions of atomic hydrogen are similar to the results obtained by Kockarts and Nicolet (1962, 1963). In the numerical computation eddy diffusion has been neglected ($\Lambda = 0$).

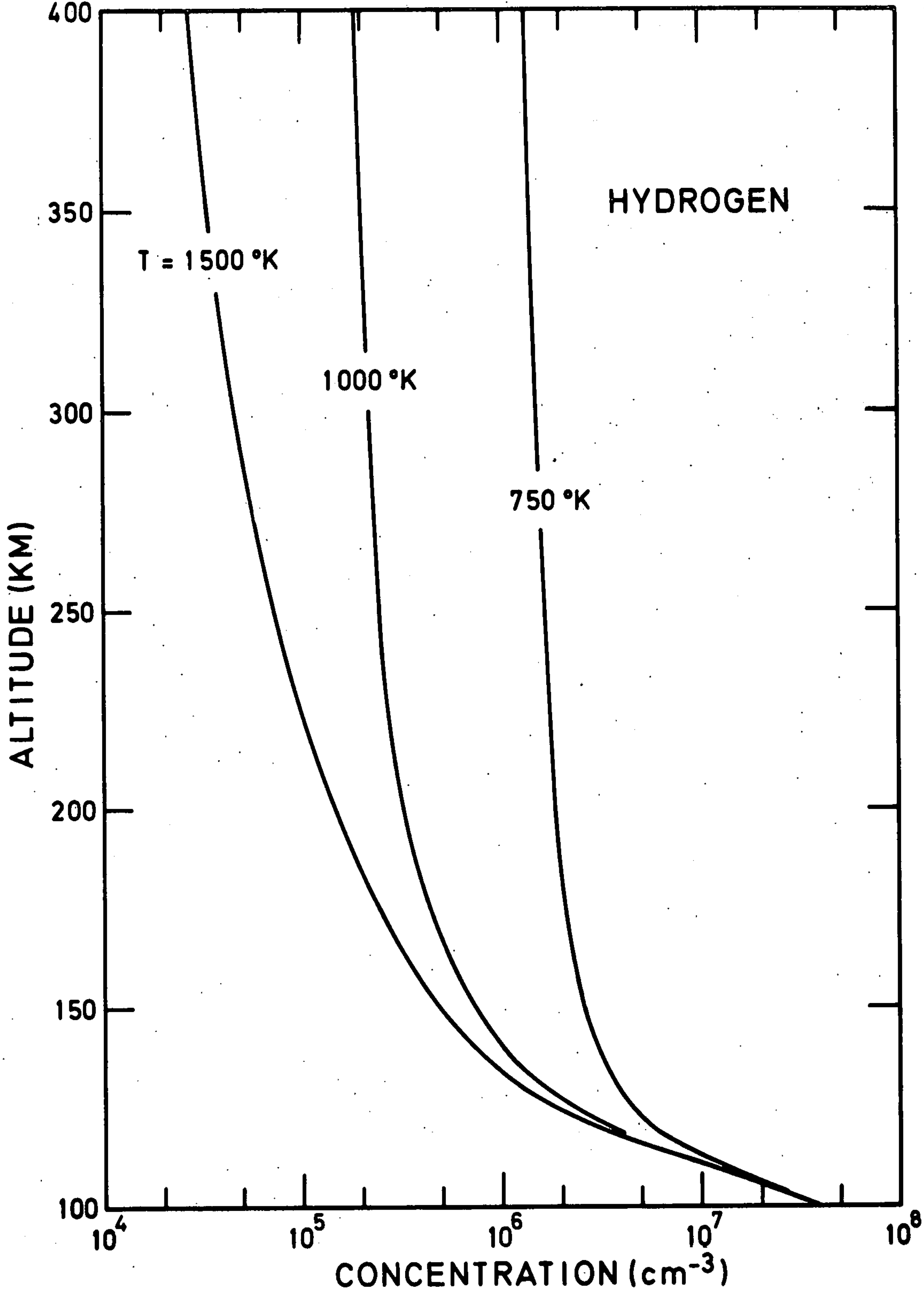

Fig. 4.- Vertical distributions of atomic hydrogen for 750°K, 1000°K and 1500°K thermopause temperatures. The concentration at 100 km is 4×10^7 cm^{-3}.

This assumption is justified by the fact that between 100 km and 120 km the vertical distribution is almost identical to a mixing distribution, even through a molecular diffusion equation has been used. The physical reason for such a situation is the presence of a real flow which is nearly identical to the maximum molecular diffusion flow. For temperatures below 1000°K, the eddy diffusion can however affect the results since upwards transport fluxes are slightly reduced.

In the case of helium, the physical situation changes since the real flux of the order of 2×10^6 cm^{-2} sec^{-1} is small compared to the maximum molecular diffusion flow which is about 10^8 cm^{-2} sec^{-1} at 100 km. When eddy diffusion is neglected, it is important to fix a level for the beginning of diffusive equilibrium. Kockarts and Nicolet (1962) have shown that an increase of 5 km for the beginning of the molecular diffusion leads to a decrease by a factor of two in the helium concentration at greater heights. Actually, there is a gradual change from mixing conditions to diffusion equilibrium. In order to analyze such a transition it is worthwhile to develop the equilibrium solution n_{eq} used in equation (14). The integration of equation (13) knowing F/D_{12} gives the expression

$$n_{eq}(z)/n_1(z_o) = (T_o/T) \exp - \left[\int_{z_o}^{z} \left(\frac{1}{H_1} + \frac{1}{H} \right) (1 + \Lambda)^{-1} \, dz' + \int_{z_o}^{z} \alpha \, (1 + \Lambda)^{-1} \frac{dT}{T} \right] \quad (22)$$

where $n_1(z_o)$ is the concentration at a height z_o characterized by a temperature T_o. If $\Lambda = 0$, equation (22) reduces to the usual equation for computing concentration in diffusive equilibrium.

The eddy diffusion coefficients are not well known at the present time. Therefore, the vertical distributions for helium, shown in Fig. 5, are computed for several height independent eddy diffusion coefficients. Since the molecular diffusion coefficient is 2.7×10^7 cm^2 sec^{-1}

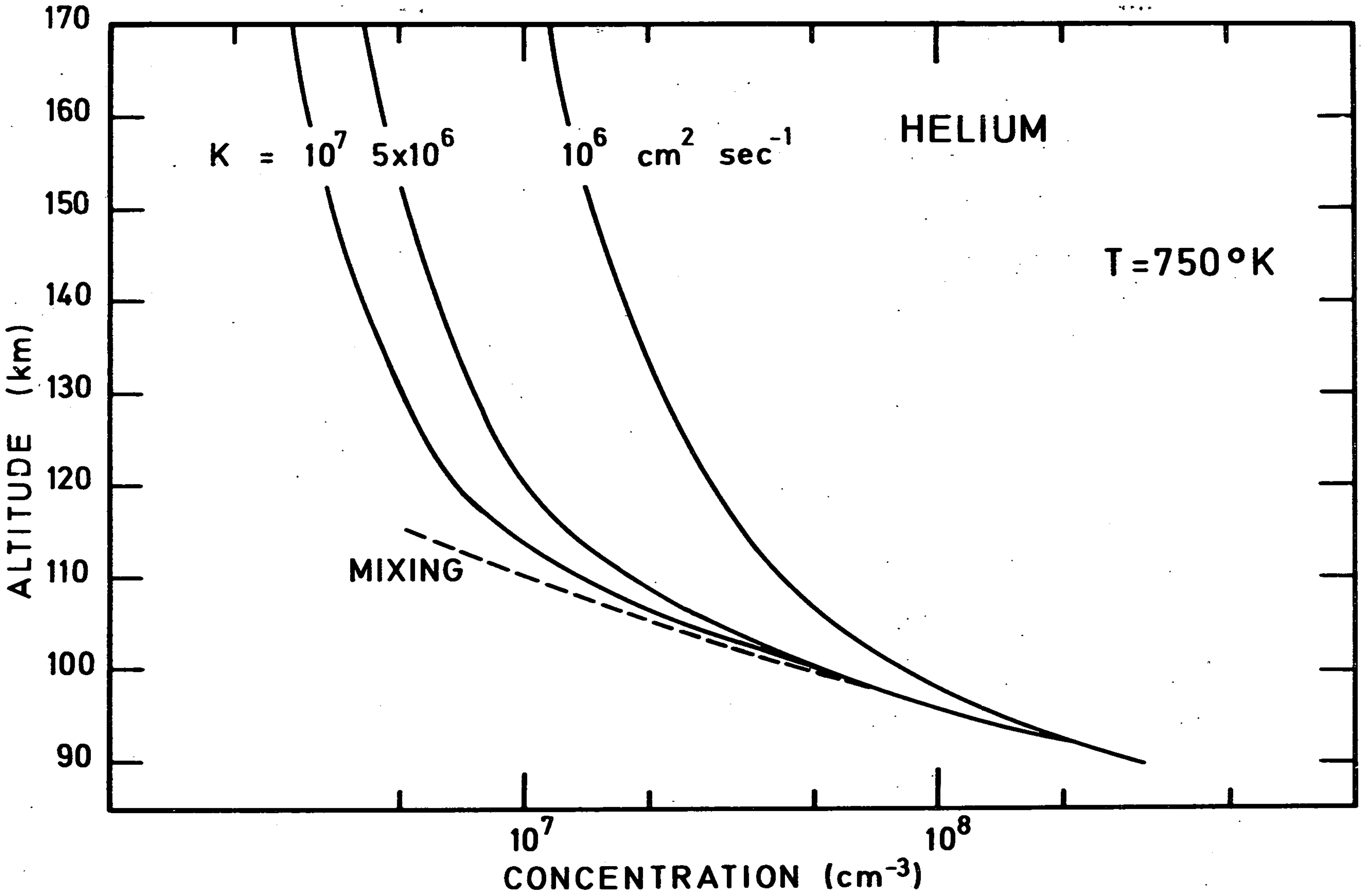

Fig. 5.- Vertical distribution of helium in the transition region for eddy diffusion coefficients of 10^7 cm^2 sec^{-1}, 5 x 10^6 cm^2 sec^{-1} and 10^6 cm^2 sec^{-1}, respectively. A complete mixing distribution is shown for comparison.

at 120 km, the ratio $\Lambda = K/D$ varies between 3.7×10^{-1} and 3.7×10^{-2} when K changes from 10^{7} cm^{2} sec^{-1} to 10^{6} cm^{2} sec^{-1} at 120 km altitude. Such a variation of K produces a change by a factor 4 in the helium concentration. It should be noted that the introduction of an eddy diffusion coefficient leads to steeper helium concentration gradients between 100 km and 120 km altitude. Those gradients are not produced by a transport flow, since equation (22) implies a zero flow condition. Fig. 6 shows how the helium concentration varies at 120 km as a function of the eddy diffusion coefficient. For a fixed thermopause temperature, it is therefore possible to explain variations in the helium concentration at greater heights by a change in the eddy diffusion coefficient in the 100 km region. However, for all the K values considered here, helium is in diffusive equilibrium above 120 km altitude. When an atmospheric model is calculated with $n(He) = (2 \pm 1) \times 10^{8}$ cm^{-3} at 120 km, this implies that below 120 km the eddy diffusion coefficient ranges between 10^{6} and 5×10^{6} cm^{2} sec^{-1}.

In order to emphazise the importance of a study of the minor constituents H and He, Fig. 7 shows how those components become major constituents above 500 km height. The difference in the extent of the helium belt in this figure compared to a similar figure given by Kockarts and Nicolet (1963) is due to an increase by a factor 4 of the atomic hydrogen concentration. Fig. 7 actually indicates the importance of diffusion in the upper atmosphere where successively an atomic oxygen belt, a helium belt and an atomic hydrogen belt are established by diffusive separation.

5.- CONCLUSION

The analysis of the general diffusion equation indicates the importance of transport processes in the Earth's heterosphere. The structure of the upper thermosphere depends strongly on the behavior of minor light constituents such as H and He in the lower thermosphere. Since the atomic hydrogen vertical flow leads to an important term in the diffusion equation,

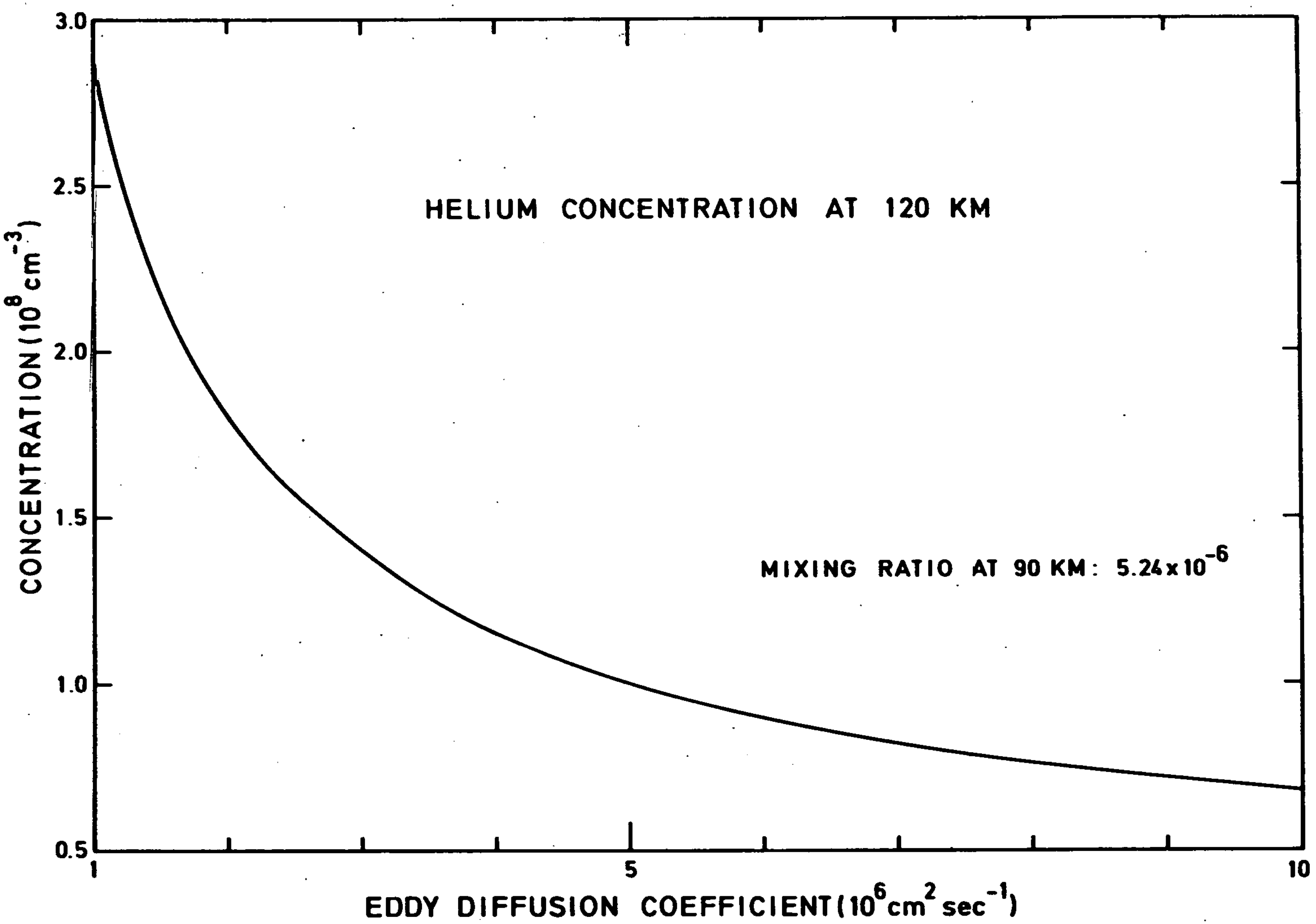

Fig. 6.- Helium concentration at 120 km as a function of the eddy diffusion coefficient adopted above 90 km.

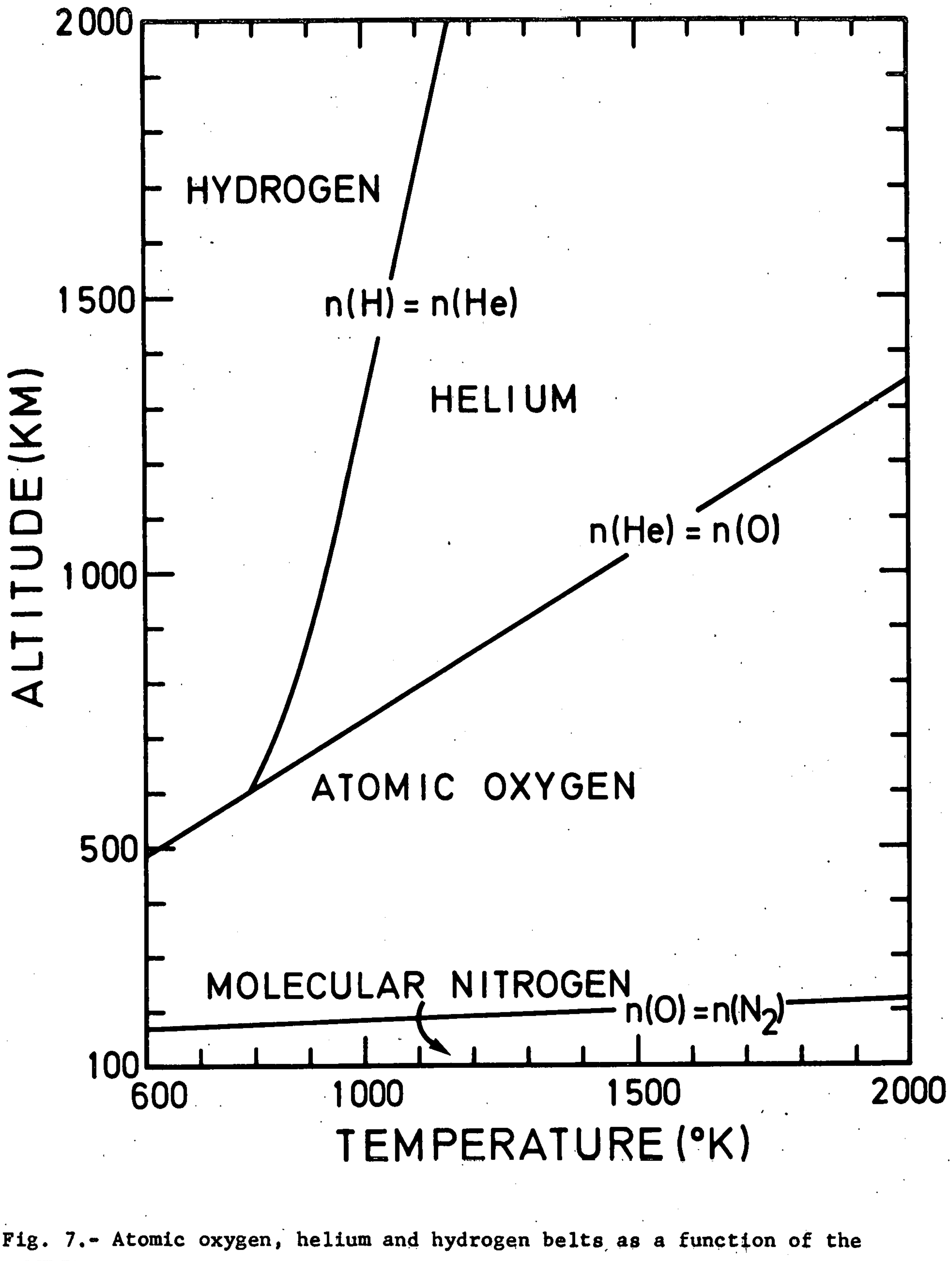

Fig. 7.- Atomic oxygen, helium and hydrogen belts as a function of the thermopause temperature. The curves give the altitude where the respective concentrations become equal.

the vertical distribution of atomic hydrogen is completely different from the vertical distributions of the other constituents which are in diffusive equilibrium. The atomic hydrogen concentration, at a fixed level in the upper atmosphere, decreases with increasing temperature. It has also been shown that eddy diffusion is not as important for atomic hydrogen as for helium. This fact is justified by the greater transport flow for atomic hydrogen. It appears that important variations of the helium concentration at 120 km can be accounted for by changes in the eddy diffusion coefficient in the neighbourhood of 100 km.

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