INSTITUT D'AERONOMIE SPATIALE DE BELGIQUE

3, avenue Circulaire, UCCLE - BRUXELLES 18

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La structure de l'atmosphère supérieure

par M. NICOLET

BELGISCH INSTITUUT VOOR RUIMTE-AERONOMIE

3, Ringlaan, UKKEL - BRUSSEL 18

VOORWOORD

De tekst van deze Aeronomica Acta is bedoeld als een hoofdstuk van het boek "Upper Atmosphere" uitgegeven door Ratcliffe, Academic Press,New York. Dit boek zal in de loop van 1960 verschijnen. De hier voorgestelde struktuur van de hogere atmosfeer heeft de verklaring mogelijk gemaakt van de dichtheden der thermosfeer, zoals ze uit de waarnemingen met satellieten in de loop van 1959 konden afgeleid worden. De bevindingen welke o.a. op enorme veranderingen in de hogere atmosfeer wijzen, zoals b.v. een dagelijkse temperatuurschommeling van 500° K, zullen het voorwerp uitmaken van een eerstvolgende Aeronomica Acta.

VORWORT

Der Text dieses Aeronomica Acta wurde geschrieben als ein Kapitel des Buches "Upper Atmosphere" herausgegeben durch Ratcliffe, Academic Press, New York. Dieses Buch wird im Laufe des Jahres 1960 erscheinen werden. Mit Hilfe der Struktur der höheren Atmosphäre, die wir vorstellen, konnte die aus den 1959 Satellitenbeobachtungen herzuleitenden Dichten der Thermosphäre erklärt werden. Die Ergebnisse, die grosse Veränderungen in der höheren Atmosphäre zeigen bzw. eine tägliche Variation von 500° K in der Temperatur, werden in ein volgendes Aeronomica Acta vorgestellt werden.

FOREWORD

The text of this Aeronomica Acta has been written as a chapter of the book "Upper Atmosphere" edited by Ratcliffe, Academic Press, New York. The book will be published in the course of 1960. The structure of the upper atmosphere presented here made it possible to interpret the densities in the thermosphere deduced from satellite observations made during 1959. The results indicating for example the tremendous variations in the upper atmosphere, for instance a diurnal temperature variation of 500° K, will be presented in a future Aeronomica Acta.

AVANT-PROPOS

Le texte de cet Aeronomica Acta a été rédigé pour constituer un chapitre du livre "Upper Atmosphere" édité par Ratcliffe, Academic Press, New York. Ce livre paraîtra au cours de 1960. La structure de la haute atmosphère que nous présentons ici a permis l'interprétation des densités de la thermosphère déduites des observations des satellites effectuées au cours de 1959. Les résultats indiquant, entre autres choses, les variations énormes de la haute atmosphère, par exemple une variation diurne de la température de 500° K, seront l'objet d'un prochain Aeronomica Acta.

THE PROPERTIES AND CONSTITUTION OF THE UPPER ATMOSPHERE

1.1. The Atmospheric Regions -

In a general description of its physical constitution and chemical composition, the Earth's atmosphere may be considered as a perfect gas composed of molecules and atoms of which only a small fraction is represented by charged particles, ions and electrons, influenced by the geomagnetic field.

However, it is not possible to determine the behavior of the neutral particles, and finally to obtain a physical picture of the upper atmosphere, without determining whether vertical distribution depends on mixing or diffusion, or on a chemical or photochemical equilibrium. In particular, it is extremely important to determine how dissociation and recombination of molecular and atomic oxygen and nitrogen are distributed with height. Furthermore, the structure of the upper atmosphere deduced from pressure or density measurements is related to the variations of the mean molecular mass depending on diffusion effects. In addition, it is necessary to know which are the most important processes of heat propagation, that is to say, convection, conduction or radiation.

Before describing the atmosphere above 50 km it is best to summarize our knowledge of the regions below (Fig. 1.).

The lower region, called the troposphere (tropos = turn or change), can be defined, in a static atmosphere, by the pressure and the temperature if the molecular mass is assumed to be constant. Meteorological soundings have shown that the temperature decreases with altitude in the troposphere reaching a temperature of about 220°K in the polar regions and about 190°K at the equator. These lower limits of the temperature are representative of the tropopause (pausis = end, cessation). The latter, however, varies in height as a function of latitude, from below 10 km in the polar regions to more than 15 km in the equatorial belt. The variations in the level of the tropopause (about 5 km) in the middle latitudes show that the changes in its structure are dependent on such atmospheric phenomena as high and low pressures. One of the essential characters of the troposphere is that the geographic equator is, to a certain extent, a " barrier" differentiating and separating the northern and southern hemispheres.

The stratosphere is essentially that region where the temperature increases, or at least does not decrease, with altitude. The stratosphere[+] extends from the tropopause to an altitude of about 50 km (stratopause[+]) where the temperature reaches a peak

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(+) Because of possible ambiguity in the terminology, S. Chapman found it preferable to use the word stratosphere to mean not the whole region above the troposphere, but only a layer between the tropopause and an upper boundary called the stratopause. We propose that the stratosphere represents a region differing from the regions below (troposphere) and above (mesosphere), and in which there are positive temperature gradients in contrast to the negative gradients in the neighboring regions.

of the order of 270°K. The seasonal variation in atmospheric ozone, and the fall-out of radioactive products artificially introduced into the stratosphere, show that, if the mass exchange in the stratosphere is sufficient to maintain mixing, the transport is certainly carried out from each side of the equator. In other words, one of the characteristic properties of the stratosphere is that meridional transport must be added to zonal transport. Since the eruption of Krakatoa in 1883, it has been known that the mean speed of deplacement around the Earth is about 120 km per hour. In addition the spread of matter northwards or southwards indicates that the meridional circulation in the stratosphere is different from that of the troposphere.

The *mesosphere* (mesos = middle) is a region situated above the stratopause (50 ± 5 km) where the temperature decreases rapidly with altitude. The temperature may reach a minimum, as low as 150 ± 10°K, at about 85 ± 5 km where the mesosphere ends at a level called the mesopause. Photochemical action is very important in the mesosphere; the chemical reactions affecting minor constituents in the air play an important role in determining atmospheric behavior. As a result of photochemical actions, airglow emissions are observed in the mesosphere. As a result not only of atmospheric motions, but also of atomic and molecular processes, the mesopause must have a tendency toward variation in height just as have the tropopause and the stratopause. The mesosphere is particularly important for the study of aeronomic processes because in it one may still follow fluctuations arising from atomic and molecular processes which manifest themselves as day-to-day variations in the hydrodynamic variables.

The essential feature of these three lower regions is that the hydrodynamical problems are practically identical. The differences are made clear at the limits. Thus the conditions in the troposphere are dependent upon the conditions at the surface of the Earth. In the heat budget of the stratosphere the effects of radiation have to be added to those of convection. Nevertheless, the characteristic common to these three regions is that the mean molecular mass of the principal constituents of the air remains constant which implies that mixing is always maintained. Because the composition remains the same, it is said that the homosphere (homos = same) extends from ground level to the mesopause.

The vertical distribution of ozone makes it possible to understand the different roles of these three regions of the homosphere. In the mesosphere, atmospheric ozone is less abundant than atomic oxygen, in an atmosphere illuminated by the sun, and there is a considerable variation between day and night. In the stratosphere, ozone whose presence depends on photochemical reactions, has a very long life and as its residence time is of the order of ten years its seasonal variations are due to meridional transport. Finally, in the troposphere, ozone is a permanent element of the air mass and its variations are intrinsically associated with the advective and dynamic transports of air. Thus, the ozonosphere is a term which cannot be associated with a definitive altitude region.

Above the mesopause there is a change in the composition and the structure of the air concurrent with the large height gradient of temperature. Because the composition of the atmosphere changes with height this region+ is called the heterosphere (heteros = other). This variation is due, first, to the partial dissociation of oxygen and, second, to diffusion.

If, however, the temperature, which is an important parameter in this region, is considered, then this region above the mesopause is called the thermosphere, because the temperature

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+ The heterosphere may begin at the mesopause. However, it is not yet possible to delimitate the exact altitude at which the mean molecular mass is sufficiently affected by the dissociation of O_2 and diffusion, and in any case this altitude can be expected to be variable due to latitudinal and seasonal effects.

in the thermosphere increases with altitude. The height at which this temperature gradient should cease could be called the thermopause, but, as we shall see later, it is possible to state that there is no thermopause if there is heat flow by conduction from the solar corona.

This height gradient of temperature in the thermosphere results from absorbed energy. Above the mesopause all ultraviolet radiation with a wavelength less than 1750 Å is gradually absorbed and an important fraction of the energy obtained may be used for heating the thermosphere. As there is only one possible means of transfer of energy by radiation [the line at 63 μ of atomic oxygen (see 4.)], the principal processes of heat transport are convection in the lower thermosphere and conduction in the upper thermosphere.

Above a certain altitude simple laws deduced from the statical equation and the equation of state (see 1.2) no longer apply because dynamic processes modify the statical picture. In this region, the <u>metasphere</u> (meta = between), the behavior cannot be explained by a static representation although the medium is still mainly unionized.

The outermost part of the atmosphere, which is almost fully ionized and where protons are more abundant than neutral hydrogen, should be called the protosphere[+]. The satellites have played an important part in the exploration of these regions, which were originally called the exosphere (exo = outside) because it was thought that in this region the laws of gas kinetics no longer applied.

To a certain extent, the study of the exosphere was concerned with the escape of neutral particles and the level at which this escape began (the beginning of the exosphere). However, the processes affecting charged particles are involved in the dynamics of the upper atmosphere which should be studied in relation to the geomagnetic field.

For several years we have placed limits on the ionosphere (that region where ions and electrons play a role). Conventional definitions of the ionosphere include that part of the atmosphere which comprises the D, E and F regions, that is to say roughly the atmospheric region between 80 and 400 km. Even in certain old and unsatisfactory systems of nomenclature the word ionosphere was used to represent an atmospheric layer between other layers, such as the layer between the stratosphere and the mesosphere. Today, following observations obtained from

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[+] According to a private communication from S. Chapman ; Shlovsky has suggested that the metasphere and photosphere should be called the geocorona.

whistlers, rockets and satellites, this can no longer be accepted because the total electron content of the atmosphere is several times the electron content below the F_2 peak. Thus the term ionosphere covers several atmospheric regions. Furthermore the fact that the word ionosphere is associated with the properties of charged particles requires that the geomagnetic field be taken into account. In addition, the dynamic processes are more important in the ionosphere than in that part of the atmosphere which is neutral. Finally, there is an association of the ionospheric phenomena with the geographic distribution of the aurora, and for this reason it is possible to make the approriate distinctions when using the terminology of Chapman. Thus, the auroral regions include those of geomagnetic latitudes greater than 60°; there is a northern auroral and a southern auroral region. The subauroral belts lie between geomagnetic latitudes 60° and 45° and are the geographical regions where auroras are occasionally seen in mean latitudes, especially near and during a sunspot maximum. In the same way, it is possible to consider a minauroral belt lying between geomagnetic latitude 45°N and S where only exceptional auroras are visible. In the analysis of geomagnetic and ionospheric observations one also speaks of the equatorial belt situated within the minauroral belt and lying between geomagnetic latitudes 20°N and S.

In finishing this brief description it should be said that in aeronomy, where particular importance is attached to the study of physical and chemical properties, and where dissociation and ionization play a primary role, the <u>minor</u> constituents should be considered to be of the same importance as the major constituents, nitrogen and oxygen.

1.2 Inter-Relation between Pressure, Temperature and Composition.

In the air the height (z) distribution of the atmospheric pressure is obtained from the differential equation relating the density ρ to the hydrostatic pressure p

$$dp/dz = - g\rho \tag{1.1}$$

where g denotes the acceleration of gravity. The density ρ is a function of the molecular masses of the various constituents m_i, and the number densities or concentration of the constituents n_i. Thus,

$$\rho = \sum n_i m_i \equiv nm \tag{1.2}$$

with the mean molecular mass defined by

$$m = \sum n_i m_i / \sum n_i \tag{1.3}$$

and the total concentration by

$$n = \sum n_i \tag{1.4}$$

The statical equation (1.1) indicates that there is no variation in height and in time of the mean mass transport velocity. If there is some variation a more complicated equation, i.e. the conservation equation of momentum, must be considered and generally does not correspond to a steady state.

Since air behaves like a perfect gas the equation of state is written as follows :

$$p = n\,k\,T \tag{1.5}$$

in which $k = 1.38 \times 10^{-16}$ erg $\deg^{-1}$ denotes Boltzmann's constant and T is the absolute temperature.

On combining the equation of state (1.5) with the statical equation (1.1) the general law governing a neutral atmosphere is obtained :

$$dp/p = dn/n + dT/T = - dz/H \qquad (1.6)$$

where

$$H = kT/mg \qquad (1.7)$$

is a quantity having the dimension of length that Chapman has termed the (local) scale height.

Since the scale height depends on three parameters (temperature, mean molecular mass and gravitational acceleration) the inter-relation between pressure, temperature and composition will depend upon the height variation of these three parameters. The variation of g is given by the known law of gravitation,

$$\frac{g}{g_0} = \left(\frac{R + z_0}{R + z}\right)^2 \qquad (1.8)$$

where g denotes the acceleration of gravity at an altitude z, g_0 the acceleration at $z = z_0$ and R + z is the appropriate distance from the earth's center. For example, g being about 980 cm sec^{-2} at ground level, g is about 950 at 100 km, 900 at 280 km, 850 at 475 km and 800 near 700 km. To a first approximation, when computations are made for a sufficiently thin layer, it is possible to use a mean value of g. If the mean molecular mass m remains constant, as in the homosphere, there is a direct relation between the pressure and the temperature. Thus, in an atmosphere with perfect mixing, a knowledge of how the pressure varies with height leads to a knowledge of the vertical distribution of the temperature and also of the concentrations.

Considering sufficiently thin layers where g and the temperature gradient are kept constant, it is possible to write

$$dH/dz = \beta = \text{constant} \tag{1.9}$$

and therefore, by integration of (1.6)

$$p/p_o = (H/H_o)^{-1/\beta} \tag{1.10}$$

and

$$n/n_o = (H/H_o)^{-(1+\beta)/\beta} \tag{1.11}$$

When the layer is isothermal, (1.10) and (1.11) can be written

$$p/p_o = n/n_o = e^{-z/H} \tag{1.12}$$

which expresses the conventional exponential decrease of pressures and concentrations.

The vertical distribution of the temperature in the whole homosphere has been determined from direct measurements of the height distribution of pressure made by means of rockets. In the heterosphere the temperature cannot be deduced direct from measurements of pressure since there is a simultaneous variation of the mean molecular mass. Since this variation is due to a dissociation effect, there is a decrease of the mean molecular mass so that temperatures deduced on the assumption of constant mass would lead to too high temperatures. On the other hand, any hypothesis about how the molecular mass varies would imply a particular vertical distribution of the temperature.

When there is a steady state of diffusion, the formula (1.6) must be written,

$$dp_i/p_i = dn_i/n_i + dT/T = -dz/H_i \tag{1.13}$$

to indicate that each constituent has a concentration or pressure gradient associated with its own mass. However, the vertical distribution in a diffusion equilibrium depends on the thermal diffusion (+) and equation (1.13) should be acceptable as a first approximation for almost all aeronomic purposes. (+) Integration of (1.13) to give

$$n_i/n_{i,o} = (H_i/H_{i,o})^{-(1 + \beta_i)/\beta_i} \tag{1.11a}$$

shows that each constituent is vertically distributed according to its own scale height gradient β_i. In any atmosphere in diffusive equilibrium, the height variation of one constituent (n_i, m_i) is known directly in terms of the corresponding quantities from another constituent (n_a, m_a) since one can write

$$\beta_i m_i = \beta_a m_a \tag{1.14}$$

(+) It is necessary to add to unity a thermal diffusion factor which in certain circumstances may be of the order of 0.3, i.e. atomic hydrogen and helium, but, it can be neglected for oxygen and nitrogen. See, K.E. Grew and T.L. Ibbs, Thermal diffusion of gases, University Press, Cambridge 1952.

It follows that an analysis of the pressure data can lead to a determination of the temperature when there is a complete mixing which keeps the ratios of all constituents constant. Any diffusion large enough to modify the value of the mean molecular mass precludes the possibility of using pressure(+) data to deduce an exact vertical distribution of temperature. Furthermore, states of simple mixing and diffusion are not the only possible ones. Suppose, in a system in either of these two steady states, a constituent (in fact, there will be oxygen) undergoes dissociation in a binary or ternary mixture (in fact, atomic and molecular oxygen and molecular nitrogen), there would then be a possibility of a change of the mean molecular mass which would modify in a very complicated manner the inter-relation between pressure, temperature and composition. It has been shown (1.4) that such a situation does exist in the thermosphere where transition from molecular oxygen to atomic oxygen occurs. In photochemical equilibrium conditions, (1.11a) is replaced by

$$n_i/n_{io} = (H/H_o)^{-3(1 + \beta)/\beta} \qquad (1.11b)$$

where β is the gradient of the atmospheric scale height.

(+) At the time of writing, no measurements of pressure has been made above 120 km. Only measurement of density have been published for regions above 140 km.

The several distributions (1.11) may be represented by means of a convenient parameter, X, which is called the vertical distribution factor. Thus, if n_i refers to a minor constituent, it may be written in terms of the mixed main atmosphere according to the representation

$$n_i/n_{io} = (H/H_o)^{-X\ (1+\beta)/\beta} \tag{1.15}$$

where X has the following values :

Mixing		1
Diffusion	$\left[(m_i/m) + \beta\right] / \left[(1 + \beta)\right]$	
Photochemical Equilibrium :		
Two-body		2
Three-body (constant temperature)		3
(varying temperature)		$3 - \beta/2\ (1+\beta)$

The vertical distribution factor clearly indicates that the greater the disparity between the molecular masses of a constituent and the atmosphere the more the diffusion equilibrium distribution deviates from the mixing distribution. The temperature variation is seen to play also a role in fixing the various concentration distributions.

1.3 Composition of the homosphere

In the study of aeronomic problems, the application of theoretical considerations will be greatly simplified in considering a subdivision between major and minor constituents. Nitrogen (about 78 °/₀) and oxygen (about 21 °/₀) are the major constituents of the dry air even if argon and CO_2 represent 0.9 °/₀ and 0.03 °/₀ of the mean molecular mass of the air, respectively.

Data for the permanent constituents of the air are listed in Table I. From this table, it is easy to deduce that in perfect mixing the constant ratio of oxygen and nitrogen concentration is

$$n(O_2)/n(N_2) = 0.2683 \qquad (1.16)$$

and, when there is a dissociation of oxygen,

$$\frac{n(O) + 2\,n(O_2)}{n(N_2)} = 0.5365 \qquad (1.17)$$

which, for the law of vertical distribution of the mass density, leads to

$$\rho = 1.34\; n(N_2)m(N_2) \qquad (1.18)$$

Since the molecular mass of molecular nitrogen M = 28.02 almost corresponds to the mean molecular mass of air M = 28.97, the vertical distribution of N_2 is almost the same in any atmosphere subject to mixing or to diffusion. A knowledge of the concentration of molecular nitrogen should therefore give fairly good estimates of the atmospheric temperature even at high levels.

Among the minor constituents, it is also useful to distinguish between the inert gases and chemically active gases in order to discuss separately the constituents which are involved in chemical or photochemical reactions. Data for the inert gases are given in Table II from which it is easy to see that the abundance ratios differ considerably from the cosmic abundance ratios.

The low concentrations of helium are due to its continuous escape at the top of the earth's atmosphere. It has been shown that the generation of alpha particles by the disintegration of uranium and thorium in the mantle and crust of the earth is of the order of one or two million particles per second and per cm^2 of surface.

Considering the present amount of helium in the earth's atmosphere (see Table II), it is easy to deduce[1] that only two million years are necessary to produce this total(+). It is

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(+) The total generation rate of α particles is $1.75 \times 10^6 cm^{-2} sec^{-1}$ Therefore, 1.13×10^{20} He^4 atoms $-cm^{-2}$ are produced in about 2,100,000 years.

appropriate, therefore, to ask how escape processes have appeared which transported helium from ground level to the outside of the atmosphere since the formation of the earth more than four billions years ago. If the escape velocity is attributed, as in the kinetic theory, to the peculiar speeds related to the temperature, it may be deduced[1] that at high levels in the thermosphere the kinetic temperature cannot be less than 2000°K.

Most of the argon, which is not the normal argon of mass M = 36 but is an isotope of mass M = 40, is of radiogenic origin . It is a result of the transformation of radioactive potassium, which is more than sufficient to have produced the present amount of atmospheric argon since the beginning of the formation of the earth. Argon being a gas which is simultaneously an inert, permanent and minor constituent of the atmosphere, is a perfect tracer for the study of diffusion, and particularly for the detection of the beginning of diffusion. For that purpose, the measurement of the ratio between the concentrations of argon and molecular nitrogen is certainly the best procedure since, as was pointed out earlier, the molecular mass of N_2 does not differ very much from the mean molecular mass of the air.

As far as the minor constituent molecules are concerned, it is difficult to be certain that no variation occurs. Several of these were discovered by spectroscopic identification of absorption bands in the infrared solar spectrum, but no systematic study of them has been made at high levels. Several others are essentially variable as shown in Table III, but all of them present variations with height. The height distributions of all these molecules differ from those of the main atmospheric constituents partly because changes occur due to chemical and photochemical actions, and partly because changes may also occur in the physical state of substances such as water. A chemical reaction is certainly the cause of the variation of ozone at ground level while in the mesosphere the principal reaction is a photochemical one.

Atmospheric methane (CH_4) probably plays a part in the production of atomic hydrogen in the thermosphere even if H_2O is involved in the photochemistry of the mesosphere. CH_4 has a biological origin since C^{14} has the normal percentage present in a biological product [2]. It is oxidised in the stratosphere and in the mesosphere and according to Bates[3], it is possible to consider oxidation rates of the order of 10^{-6} per second at the stratopause level. In order to replace the dissociated methane, an upward transport of CH_4 from the troposphere to the stratosphere must therefore occur and the flow is subject to the law of exchange of mass between these two atmospheric regions. Thus, atomic hydrogen and its various compounds should have in the mesosphere a value of the order 5×10^{-6} of the main atmospheric gases, if there is a steady state.

The loss rate of CH_4 could be as large as 10^{10} to 10^{11} molecules $cm^{-2}sec^{-1}$ which would imply that the whole atmospheric methane content would be replaced in about 10 to 100 years. In any event, it is certain that atomic hydrogen must exist in the lower thermosphere in quantities which depend on processes affecting methane and water vapor in the mesosphere.

If the vertical distribution of N_2O above the tropopause followed the same atmospheric distribution as in the troposphere, about 10^{11} N_2O molecules $cm^{-2}sec^{-1}$ would be dissociated. Since the association of N_2 and O is a very slow process, depending on three body collisions, there is a deflection with height of the concentration of that molecule which can be neglected at mesospheric levels.

On the other hand, nitric oxide and nitrogen dioxide which are not generally considered in the lower atmosphere are produced in the mesosphere and thermosphere.

Twenty five years ago it was generally accepted that the composition of the stratospheric air varied with altitude, because helium and oxygen samples taken from heights above 20 km showed departures from a mixing distribution toward partial diffusive separation. Comparison of oxygen and nitrogen percentages in the stratosphere showed that the relative amount of oxygen decreased with height. Paneth[4] has suggested as a probable explanation that oxygen overreacts with collecting chambers. Although early helium analysis showed an increase in the relative content in the stratosphere (16 to 25 km), Gluckauf and Paneth[5] now believe that the percentage of helium remains constant in the stratosphere just as it does in the troposphere. A relative upwards increase of helium and neon and a decrease of argon have been deduced from the analysis of rocket samplings in the mesosphere from 60 km to 90 km. This does not however necessarily imply an effect of diffusive separation but may be explained by mass discrimination in the sampling system(+). First mass spectrometric observations[6] taken at altitudes up to the thermosphere (137 km) showed there was practically no change in the ratio of argon to molecular nitrogen. However, more recent observations by Meadows and Towsend[7] in the winter night time show that diffusive separation of argon and nitrogen exists in the region 110 to 150 km and may begin at about 105 km. Thus, it is possible to conclude that gases which are sufficiently inert chemically do not exhibit changes in relative abundance below 100 km.

Friedman and his collaborators[8] have used photon counters in a series of rocket flights to measure solar radiation near 1500 A. These measurements showed how the variation of the

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(+) A fuller account will be found in 'Rocket Exploration of the Upper Atmosphere' (R.L.F. Boyd and M.J. Seaton, ed.) Pergamon Press, London, 1954.

molecular oxygen absorption varied with height and consequently the vertical distribution of O_2 between 110 and 180 km is known. The results show that partial dissociation of O_2 occurs above 100 km, but it is possible to show that the observed concentrations are larger than those deduced from photochemical equilibrium and in fact correspond to an upward diffusive transport of molecular oxygen. This point will be discussed in detail later with the general subject of diffusion.

In conclusion, information on the composition of the atmosphere up to 100 km indicates that the homosphere extends certainly to the mesopause. A constant ratio of molecular oxygen and nitrogen concentrations is maintained by mixing. Beyond the mesopause, that is to say in the thermosphere, a varying mean molecular mass must be taken into account even though there is still mixing, for partial dissociation of oxygen must occur. Nevertheless, the effect of mixing will decrease with height compared to the increasing power of diffusive separation. Finally, the mean of the molecular mass will decrease in the thermosphere according to the vertical distribution of oxygen and nitrogen in diffusive equilibrium.

1.4. The Dissociation of molecular oxygen

Since Chapman's study(+) in 1930, it has been supposed that oxygen dissociates at a height of about 100 km, and in all subsequent theories(++) the problem was attacked from the standpoint of an assumed photochemical equilibrium where the photodissociation and recombination processes are considered to yield the same number of processes $cm^{-3}sec^{-1}$ (statistical equilibrium) or the same energy ($erg\ cm^{-3}sec^{-1}$, radiative equilibrium). In other words, the times of dissociation (τ_{diss}) or of recombination (τ_{rec}) were assumed to be small compared with the time of diffusion (τ_{diff}) or the time of mixing (τ_{mix}). However, general conditions such as

$$\tau_{rec} < \tau_{mix} \tag{1.19}$$

or

$$\tau_{diss} < \tau_{diff} \tag{1.20}$$

cannot be accepted so that there are serious departures from photochemical equilibrium and the condition (1.17) cannot be

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(+) See a general account by S. Chapman, The Photochemistry of Atmospheric Oxygen, Rep. on Progress in Phys., 9, 92 (1943).

(++) For a fuller account of the various theories by M. Nicolet and P. Mange see The Dissociation of Oxygen in the High Atmosphere, J. Geophys. Res. 50, 15, (1954).

used in the problem of the transition region of oxygen dissociation.

Inasmuch as the dissociation of molecular oxygen begins to occur, for increasing height levels, in the lower thermosphere one is no longer free to assume a constant molecular mass at these levels so that, as stated before, there is no simple relation between the measured pressure and the temperature. In fact, vertical transport can carry molecular oxygen upward whilst the atomic oxygen resulting from dissociation is carried downward by diffusion. The process is as follows : one dissociated O_2 molecule is replaced by another through upward diffusion while oxygen atoms are forced downward until they reach a region in which the pressure is sufficient to cause them to recombine.

Let us consider first the general equation[9] for the photochemical and recombination effects influencing molecular oxygen,

$$dn(O_2)/dt + n(O_2)J = a\ n^2(O) \qquad (1.21)$$

in which J denotes the rate coefficient for dissociation of O_2 (per molecule sec^{-1}) and a is the recombination coefficient. At sufficiently high levels $n(O_2) > n(O)$, and any variations of $n(O_2)$ hardly affect the magnitude of n(O). When the optical depth is negligible J may be taken as a constant. Thus, the solution of (1.21) is

$$n(O_2) = n_o(O_2)e^{-Jt} + \frac{a\,n^2(O)}{J}\left[1 - e^{-Jt}\right] \qquad (1.22)$$

where $n_o(O_2)$ is the O_2 concentration at time $t = 0$.

The time necessary to reach 50 °/ₒ of the photochemical equilibrium value (or of any initial value) is called the time $\tau_{diss}(O_2)$ of dissociation of molecular oxygen and is thus

$$\tau_{diss}\,(O_2) = 0.7/J \qquad (1.23)$$

A calculation of the dissociation rate coefficient J requires the computation of the dissociation in the Schumann-Runge continuum of O_2 by solar radiation of wavelengths shorter than 1750 A. Numerical results [9] lead to a value of J of the order of $5 \times 10^{-7} \sec^{-1}$ at zero optical depth so that the dissociation time of molecular oxygen in the solar radiation field is of the order of 15 days. In the actual atmosphere with finite optical depth the times of dissociation of molecular oxygen correspond to not less than 15 days and if diffusion times are shorter than that, there is an upward vertical transport of oxygen molecules[10].

The long time required for the dissociation of molecular oxygen is due to the low radiation temperature observed on the spectral range of the Schumann-Runge continuum. In order to show how much the dissociation rate coefficient is affected by the radiation temperature, various values calculated by M. Nicolet and P. Mange[9] are presented in Table IV. The temperature dependence shows that the radiation energy is the most important parameter which leads to different conclusions. The time of dissociation is less than one hour for a radiation from a black body at 6000°K, less than one day for 5000°K, of the order of 10 days for 4500°K, and about four months for 4000°K. Thus, it is easy to understand why no departure from photochemical equilibrium was considered to be possible when radiation temperatures greater than 5000°K were adopted.

As far as atomic oxygen is concerned, the equation may be written as follows :

$$dn(O)/dt + 2\, a\, n^2(O) = 2n(O_2)J \tag{1.24}$$

During dark hours J = 0 and the solution of (1.24) is

$$n(O) = \frac{n_o(O)}{1 + 2\, a\, n_o(O)\, t} \tag{1.25}$$

so that n(O) is reduced to 50 °/o of its initial value $n_o(O)$ in a time τ_{rec} (O) given by

$$\tau_{rec}(O) = 1/2\, a\, n_o(O) \tag{1.26}$$

The value

$$a = 5 \times 10^{-34} n(M)\, T^{1/2}\ \mathrm{cm}^6 \mathrm{sec}^{-1} \tag{1.27}$$

has been deduced[11] from a theory in which molecules are formed by three-body collisions involving a third particle with concentration n(M). Such a value is obtained when the collision probability for molecular formation is about 50 °/o. In order to obtain orders of magnitude, we adopt for T an arbitrary value 400°K, and we obtain

$$\tau_{rec}(0) = \frac{5 \times 10^{3}}{n(M)n_{o}(0)} \text{ sec} \tag{1.28}$$

In Table V the life times of an oxygen atom are given corresponding to different values of the concentration of the third particle (total number density) and of the initial concentration of atomic oxygen. Since we know that at altitudes above 120 km both concentrations are certainly less than 10^{12} particles cm^{-3} the life times of oxygen atoms are more than one year. There is therefore downward transport of all oxygen produced in a sunlit atmosphere. Even at 100 km where the total concentration can reach 10^{13} cm^{-3} the recombination time is not less than one month so that even there transport processes are allowed to act and, finally, to modify the whole structure of the transition region of oxygen dissociation which would be obtained under photochemical equilibrium conditions.

The introduction of atmospheric data concerning pressure and temperature makes it possible to consider numerical magnitudes[12] and to reach interesting conclusions(+). First, the production peak of atomic oxygen, caused by photodissociation of molecular

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(+) It is not possible to consider precise values of pressure, since observational data are very different at 100 km. An error of 5 km may be possible.

oxygen in the Schumann-Runge continuum at wavelengths shorter than 1750 A, occurs between 90 and 95 km with a rate coefficient of about 1.5×10^5 atoms $cm^{-3}sec^{-1}$. The concentration peak should be observed near 110 km under conditions of photochemical equilibrium and would have a value of the order of 1.7×10^{12} atoms cm^{-3}. Such a concentration would be of the order of the nitrogen concentration and would lead to a total pressure greater than that measured at 110 km. Therefore, $n(O) = 1.7 \times 10^{12}$ cm^{-3} may be considered as the maximum value which can be reached at 110 km.

Second, the production of oxygen atoms at a rate of about 4×10^{11} oxygen atoms per $cm^2 sec$ must be balanced by the same total number of recombinations.

Writing

$$\int_z^\infty n(O_2)J \; dz = \int_z^\infty a \; n(M) \; n^2(O) \; dz \qquad (1.29)$$

it is possible, by adopting a mean value for n(O), to conclude that

$$\left(\overline{n(O)}\right)^2 a \; n(M) \; H = 2 \times 10^{11} \; cm^{-2}sec^{-1} \qquad (1.30)$$

in which H denotes the atmospheric scale height, $\overline{n(O)}$ being the concentration near the production peak (90-95 km); $\overline{n(O)}$ has a value of the order $1.7 \times 10^{12} cm^{-3}$. Concentrations obtained from (1.30), which cannot be applied above the production peak, show that the maximum concentration of a layer formed by continuous transport and for which the thickness is of the order of the scale height, varies between 7×10^{11} cm^{-3} at 85 km and 1.7×10^{12} cm^{-3} at about 95 km. Fig. 2 represents the vertical distribution of atomic and molecular oxygen in the lower thermosphere. By comparing the possible concentrations of atomic oxygen between 100 km and 110 km, it is possible to determine two extreme limits. The downward transport leads almost to

$$n(N_2) = 10\, n(O) \tag{1.31}$$

and the photo-equilibrium conditions to

$$n(O) = 2\, n(N_2) \tag{1.32}$$

instead of the constant ratio defined by (1.17) which is approximately

$$n(O) = \frac{1}{2}\, n(N_2) - 2n(O_2) \tag{1.33}$$

A constant mixing ratio (1.33) as a result of perfect mixing would lead to a concentration peak of atomic oxygen just below 90-95 km, and, therefore, a vertical distribution above that peak following the atmospheric distribution of O_2 and N_2. Such a low value for the concentration of atomic oxygen $[n(N_2) \simeq 10\, n(O_2)]$ in the lower thermosphere would prevent this atom from playing an important role except at extreme altitudes.

After considering the extreme conditions it is difficult to define the normal conditions, although it is possible to say that the concentration of atomic oxygen between 90 and 110 km is of the order of $(1.0 \pm 0.5) \times 10^{12}$ cm^{-3}. The curves in Fig. 2 show the vertical distributions of atomic oxygen for vertical transport downward to below the production peak [between curves (a) and (b)] , and for photochemical equilibrium [curve (c)].

Since the peak of atomic oxygen is situated in the lower thermosphere it is safe to conclude that the heterosphere begins above the mesopause. The mesopause may be considered as the extreme lowest border of the homopause, where a change in the mean molecular mass occurs. Because the mesopause is the level where a temperature minimum is reached, the physical conditions will vary with temperature, with changes of the order of 70°K involved. In the mesosphere, atomic oxygen is subject to photochemical and chemical actions, leading to different aeronomic conditions between day and night, such as increase of the mesospheric ozone at night, while its vertical distribution in the thermosphere depends on the downward transport and its maximum concentration is related to a balance between the total production and total recombination. Because the absolute concentration is varied by atmospheric motions, fluctuations in some aeronomic processes, such as the emission of the green line of atomic oxygen, can be understood.

1.5 The problem of diffusion.

Before discussing the effect of diffusion in the thermosphere it is useful to introduce certain velocities by which one-dimensional diffusion motions only may be described.

Suppose each molecule has a velocity c with respect to an arbitrary reference frame, then the mean velocity of the molecules $\bar{c}$, in an infinitesimally small region is obtained by summing all the individual velocities and dividing by the number n of molecules. Since $\bar{c}$ is in fact a point function and may vary in space, n $\bar{c}$ is the number of molecules per unit time passing through unit cross section of a plane at rest with respect to the reference frame. If the gas consists of several types of molecules, the mean velocity $\bar{c}$ is defined by the total particle transport at a given point through the expression

$$n\bar{c} \equiv \Sigma n_i \bar{c}_i \tag{1.34}$$

where $\bar{c}_i$ denotes the mean velocity associated with molecules of the i^{th} type. Summing the mass transport $n_i m_i \bar{c}_i$ the mean mass transport velocity c_o is defined at a given point by

$$\Sigma n_i m_i \bar{c}_i = \Sigma \rho_i c_i \equiv c_o \rho \tag{1.35}$$

The velocity $\bar{c}_i$ is not the same as $\bar{c}$, when one constituent is moving relative to the general motion of the gas at a specified point. The peculiar velocity, or the mean thermal velocity

$\bar{C}_i$ is then defined by

$$\bar{C}_i \equiv c_i - \bar{c}$$

Considering two kinds of molecules, (1.34) leads to

$$\bar{C}_i - \bar{C}_2 = \bar{C}_1 - \bar{C}_2$$

a quantity which is not altered by the reference velocity.

In a physical situation the conservation equations of mass, momentum and energy are involved. They are, following Chapman and Cowling[13]

$$\partial\rho/\partial t + \partial(\rho c_o)/\partial t = 0 \qquad (1.36)$$

$$\partial c_o/dt + c_o\,\partial c_o/dz + (1/\rho)\partial p/\partial z + g = 0 \qquad (1.37)$$

$$\partial p/dt + \partial(c_o p)/\partial z + \frac{2}{N}\,p\,\partial c_o/\partial z + \frac{2}{N}\,\partial q/\partial z = 0 \qquad (1.38)$$

where q represents the vertical component on the heat flow vector and N denotes a factor depending of the internal energy of molecules; N = 3 for monoatomic gases and N = 5 for diatomic molecules considered as rigid and elastic spheres.

For a two constituent gas q is given by

$$q = -\lambda_c\,T/z + k\,n\,T\,\alpha_T\,n_1 n_2/n^2\,(\bar{C}_1 - \bar{C}_2) + \frac{5}{2}\,k\,T(n_1\bar{C}_1 + n_2\bar{C}_2) \qquad (1.39)$$

in which λ_c is the coefficient of thermal conductivity and α_T the thermal diffusion factor, which is almost a constant.

The first term represents the thermal flux by ordinary conduction in a gas consisting of a single kind of molecule.

The second term represents the heat transport by diffusion. The mean relative velocity of the two constituents $(\bar{C}_1 - \bar{C}_2)$ is given by

$$\bar{C}_1 - \bar{C}_2 = \frac{n^2}{n_1 n_2} D \left[\frac{\partial (n_1/n_2)}{\partial z} + \frac{n_1 n_2 (m_2 - m_1)}{n \rho} \frac{\partial \log p}{\partial z} + \frac{n_1 n_2}{n^2} \alpha_T \frac{1}{T} \frac{\partial T}{\partial z} \right] \tag{1.40}$$

in which D denotes the diffusion coefficient. This term will disappear when there is no mutual diffusion, i.e. when the steady state has been obtained.

The third term is associated with the general motion of the gas, for each molecule carries an average quantity of heat energy equal to $(5/2)kT$. When this term is neglected, that is, when the mean velocity is zero, the total number of molecules per unit volume remains constant, so that $\partial n/\partial t = 0$, and molecules pass equally frequently in each direction through a plane moving with the gas. The mass velocity may, however, differ from zero, i.e. $\partial \rho/\partial t \neq 0$.

For a discussion of dynamic processes the statical equation (1.1) is modified as indicated in (1.37) to take account of diffusion effects. However, these are certainly small compared with the effects of gravitational acceleration, and the law of conservation of momentum (1.37) may be considered identical, to a good degree of approximation, with the statical law relating the mass density ρ to the hydrostatic pressure p.

The conservation of mass for each constituent is governed by the expressions

$$\partial n_1/\partial t + \partial(n_1 c_1)/\partial z = 0 \qquad (1.41)$$

and

$$\partial n_z/\partial t + \partial(n_2 c_2)/\partial z = 0 \qquad (1.42)$$

The change of pressure with time is due essentially to the mass velocity while the variation of the heat flux with height is due to the variation of the mass velocity; that is, (1.38),

$$\partial p/\partial t = - g \rho c_o \qquad (1.43)$$

and

$$\partial q/\partial t = \frac{5}{2} p \, \partial c_o/\partial z \qquad (1.44)$$

In equation (1.43) products of derivatives of first order quantities in the integration of the momentum equation (1.37) have been neglected.

If the mass velocity had been neglected, there would be (see equation 1.43) no variation of pressure. In other words, when an aeronomic problem is considered, it is necessary to know whether or not the total pressure is affected. If the constituents are major constituents, the variation of pressure with time cannot be neglected and is a direct result of diffusion. Application of the theory to atomic oxygen and molecular nitrogen shows that the pressure increases markedly when diffusion equilibrium replaces mixing equilibrium.

In the study of the behavior of a minor constituent the problem of diffusion is much simpler. In effect, the total pressure remains practically constant in a mixed atmosphere while the minor constituent diffuses upward or downward. Therefore, the following conditions can therefore be applied for a minor constituent :

(i) $$\partial p/\partial t = 0 \tag{1.45}$$

showing that the main atmospheric distribution is fixed in time ;

(ii) $$n = n_1 + n_2 \simeq n_2, \tag{1.46}$$

indicating the difference between concentrations of major and minor constituents;

(iii) $$n_1 c_1 + n_2 c_2 = 0 \tag{1.47}$$

assuming that there is no net particle flow across any surface.

Then, if the velocity c_1 of the minor constituent is denoted by w for convenience, (1.40) becomes, after using conditions (1.45) (1.46) and (1.47),

$$w = - D\left[\frac{1}{n_1}\frac{\partial n_1}{\partial z} + \frac{m_1 g}{kT} + (1 + \alpha_T)\frac{1}{T}\frac{\partial T}{\partial z}\right] \qquad (1.48)$$

Using the preceding relations (1.7) and (1.9), we write for the main constituent[14]

$$\beta/H = (1/T)\partial T/\partial z = (1/H)\partial H/\partial z \qquad (1.49)$$

$$(1+\beta)/H = - (1/n)\partial n/\partial z \qquad (1.50)$$

and for the minor constituent, (1.15),

$$X(1 + \beta)/H = -(1/n_1)\partial n_1/\partial z \qquad (1.51)$$

where X is the vertical distribution factor introduced in (1.15). When X = 1, the constituent follows a mixing distribution.

The vertical velocity w of the minor constituent given by (1.48) takes the following form, using (1.49) and (1.51) :

$$w = \frac{D}{H}\left[X\ (1+\beta) - \beta\ (1+\alpha_T) - (m_1/m)\right] \qquad (1.52)$$

The vertical velocity, w, increases almost exponentially with altitude and is upward if

$$X > X_D = \frac{\beta(1 + \alpha_T) + m_1/m}{1 + \beta} \qquad (1.53)$$

and downward if

$$X < X_D \qquad (1.54)$$

Although the direction of diffusive transport is naturally dependent on the relative mass (m_1/m) of the constituent, it must be pointed out that the concentration gradient may change the whole picture. For example, any photochemical equilibrium of molecular oxygen corresponds to (1.53) and there is a continuous upward transport. Any production of oxygen atoms in the thermosphere corresponds to (1.54) and there is a continuous downward transport.

The continuity equation of diffusion for vertical motion, according to (1.41) and (1.47), is

$$\partial n_1/\partial t = \partial(n_1 w)/\partial z \tag{1.55}$$

in which the right side is the one-dimensional counterpart of the usual divergence term. With the use of (1.52), the neglect of thermal diffusion ($\alpha_T = 0$), and an expression for the diffusion coefficient D, it is possible to explain the continuity equation (1.55). Here, we describe molecules as elastic spheres with a collision diameter σ, and D is written(+)

(+) The diffusion coefficient is proportional to $T^{1/2}/n$. In fact, the temperature exponent may be between the two extreme values 1/2 and 1, corresponding to elastic spheres and Maxwellian molecules, respectively. Furthermore, interactions which can occur between certain atoms may modify the absolute value of the coefficient. See, for example, A. Dalgarno, 'The mobilities of ions in their parent gases', Phil. Trans. Roy. Soc. London 250, 426 (1958).

$$D = (3/32n\pi\sigma^2)\ (1 + m_1/m)^{1/2}\ (8\ kT/\pi m_1)^{1/2} \qquad (1.56)$$

$(8kT/\pi m_1)^{1/2}$ is the mean molecular velocity of gas molecules of mass m_1 at the temperature T, and $(8kT/\pi m_1)^{1/2}\ (1+m_1/m)^{1/2}$ is the mean relative velocity for two constituent gases.

Thus, (1.55) is written, (1.52) and (1.56),

$$\partial n_1/\partial t + n_1 w\left[\frac{2+\beta}{2(1+\beta)} - X\right]\frac{1+\beta}{H} + n_1\frac{(1+\beta)D}{H}\ \partial X/\partial z = 0 \qquad (1.57)$$

or

$$\partial n_1/\partial t = n_1\frac{D(1+\beta)}{H^2}\left\{\left[\beta + m_1/m_2 - X\ (1+\beta)\right]\left[\frac{2+\beta}{2(1+\beta)} - X\right] - \partial X/\partial z\right\} \qquad (1.58)$$

Stationary state solutions are obtained for values of the vertical distribution factor X which are solutions of the appropriate Riccati equation. The equilibrium distribution is obtained by equating the first bracket of equation (1.58) to zero.

The solution of (1.58) has been worked out by Epstein[15] for an isothermal atmosphere and by Mange[16] for the more general case of an atmosphere with varying scale height. A still more general solution which took account of sources and sinks due to photochemical action or atomic production by meteors has also been obtained by Mange[17]. An important result of Mange's work is that diffusion times deduced directly from expressions of diffusion velocities differ from those obtained by detailed consideration of equations such as (1.58). In fact, it is possible to deduce the time necessary to change the concentration by a fixed percentage, when diffusion proceeds from an initial X state at any specified level. In fact, the times to reach diffusion equilibrium strongly depend on initial and final concentration gradients. Furthermore, the variation of the concentration at a given altitude is a function of the concentration at lower altitudes.

In order to obtain a sufficiently precise estimate of the situation, it is necessary to adopt a working model of the thermosphere in the form of a primary model in which conditions of pressure and temperature are not too far from the final and actual conditions.

Let us consider an atmosphere in which the scale height at a given reference level is $H = 10$ km and has a constant gradient $\beta = dH/dz = 0.2$. Adopting an average molecular mass $M = 24$, which lies between those of atomic and molecular oxygen, the atmospheric model is defined by taking the total concentration $n(M) = 1.25 \times 10^{12}$ cm^{-3} at $H = 10$ km .

Such an atmospheric model may correspond to an initial thermosphere in which the temperature is 273°K between 110 km and 115 km. Since $M = 24$, atomic oxygen concentration must be 4.2×10^{11}cm^{-3} and molecular nitrogen concentration 8.3×10^{11}cm^{-3}. Since at the start the mean molecular mass is constant the equivalent height can be found and an atmospheric model such as that shown in Table VI may be used for describing diffusion phenomena. It may be pointed out that slight modifications are possible to fit observational data in the region of 100-120 km, but they will not modify the whole argument.

Let us consider first the departure from conditions under which molecular oxygen is in photochemical equilibrium. Starting from conditions of photochemical equilibrium, there is an increase in the molecular oxygen concentration according to the continuity equation

$$\partial n(O_2)/\partial t = - n(O_2)J - \partial\left[n(O_2)\ w\right]/\partial z \qquad (1.59)$$

which expresses the loss of oxygen molecules by photodissociation and transport by diffusion. The recombination is neglected since it always has the same sign as the diffusion term, and is, moreover, always small compared with it.

The final result is shown in Fig. 3 in which the continuous curves represent the effect of diffusion overcoming the effect of photodissociation. It has been assumed that solar radiation is present for twelve hours each day, and it can be seen that the concentration changes are extremely rapid, several orders of magnitude in a few days in regions above 140 km.

Comparing these changes (Fig. 3) in the molecular oxygen concentrations with arbitrary diffusion equilibrium distributions, it is possible to determine the rapidity with which a particular diffusion equilibrium is reached. For example, let us assume that below H = 10 km a photochemical equilibrium can be maintained and that above this level diffusion can take place. Fig. 3 shows that for the whole thermosphere above H = 10 km a vertical distribution of oxygen in diffusive equilibrium is produced in about 2 days. Under the same conditions but with H = 8 km a distribution of O_2 in diffusive equilibrium will take less than 10 days. Since H = 8 km may represent an altitude between 100 km and 105 km photochemical equilibrium is replaced by diffusive equilibrium because the diffusion time is shorter than the dissociation time for molecular oxygen.

It is clear that no other atmospheric transport process can oppose the upward diffusion of molecular oxygen. If mixing is effective, its action will reinforce the upward motion, and may lead to a vertical distribution following the general atmosphere for which the height gradient is less steep than that for diffusion. Therefore, it is necessary to consider the other aspect of diffusion, i.e. the conditions of diffusive separation of constituents in an atmosphere in which initial conditions are represented by mixing.

The effect of diffusion on the distribution of molecular oxygen originally in mixing equilibrium is shown in Fig. 4. The different curves represent the effect of diffusion for successive days during one week.

The changes in the concentration of O_2 (M = 32), a minor constituent in an atmosphere of atomic oxygen and molecular nitrogen (M = 24), are extremely rapid. Fig. 4 shows that only one day is necessary to reach a diffusion distribution above the level where H = 12.5 km. In the atmospheric model used for the calculation (see Table IV) this corresponds to an altitude at which the total concentration would be less than $3 \times 10^{11} cm^{-3}$ or the density would be of the order of $10^{-11} gm\ cm^{-3}$. Another interesting level corresponds to H = 10 km, because this may be the level at which diffusive equilibrium begins, if diffusion is possible in three days. Fig. 3 shows how above H = 10 km, or an altitude of 110-115 km according to data of Table IV, a diffusive distribution of O_2 results after three days. If the two distributions of O_2 (Figs 3 and 4) are compared with their limits for photochemical and mixing conditions, it can be said that the diffusion conditions beginning at H = 10 km (an altitude of 110-115 km) should represent the mean of the two extreme possibilities. In other words, if the diffusion times (2 to 3 days) necessary for photochemical or mixing conditions are the same, a diffusive equilibrium represents the normal condition when dissociation and mixing times are more than 3 days.

When making measurements of the variation of the concentration of a constituent with altitude, at lower altitudes, it is necessary to specify the vertical distribution as a function of time and concentration, because the mixing time determines the lower boundary condition of the continuity equation related to the transport by diffusion. Argon compared with molecular nitrogen is such an example which can be used to determine the transition from the mixing to diffusion states.

If the working model is the same as that used for discussing the diffusion of molecular oxygen (Table IV), it is possible to obtain a sufficiently precise estimate of the situation. It must be pointed out, however, that 'diffusion times' may have different meanings according to the conditions involved.

If diffusion time is defined as the time necessary to increase concentrations by a certain factor at a certain height when diffusion is upward (helium), or decrease it by the same factor when it is downward (argon), absolute times can be defined by using the continuity equation. In that case, the shortest times are found for constituents with masses which differ greatly from the mean molecular mass, and the longest times for constituents having masses comparable with the mean molecular mass. In an aeronomic problem, however, another criterion for diffusion is necessary, one in which a direct comparison is made above a certain altitude between a mixing distribution and a diffusion distribution. In such a case, the diffusion times have a completely different meaning and are directly related to the absolute values of the concentrations in a definite atmospheric region.

Fig. 5 and Fig. 6 show the results for argon and helium, respectively. The initial vertical distributions are represented by the principal constituent of mass M = 24 and diffusion occurs for minor constituents of mass M(A) = 40 and M(He) = 4. The final states are vertical diffusion distribution corresponding to three starting levels H = 8 km, 10 km and 12.5 km.

In this way characteristic times are derived for redistribution of a minor constituent originally completely mixed, and are related to different diffusion distributions beginning at a series of different levels.

One feature of Figs 5 and 6 is immediately apparent : the rapidity with which a diffusion equilibrium is reached depends on the altitude. For example, if diffusion is not counteracted by mixing above H = 8 km (between 100-105 km) seven days (see Fig 5) are necessary to reach a diffusion equilibrium for argon in the thermosphere above that altitude. During the same time (Fig. 6) diffusion will have no important effect on helium below H = 10 km (between 110-115 km). In other words, diffuse transport produces an equilibrium distribution above H = 10 km in about 3 days for argon and about 7 days for helium. Finally, above H = 12.5 km the shortest times of diffusion are involved : less than one day is required to reach diffusion equilibrium for argon and less than 3 days for helium.

These various criteria describe clearly how concentration changes occur in the thermosphere under the continuous effect of diffusion and finally lead to a clear conclusion that above a certain level in the lower thermosphere mixing must ultimately be replaced by diffusion. If the time of mixing is introduced the discussion may be limited by arbitrary conditions, since we do not yet know the coefficient of mass exchange at altitudes of 100 km. Thanks to recent observations of Meadows and Townsend[7], it is possible to assume that the mixing times for argon are not less than one day and not more than one week between 100 km and 110 km, altitudes which are represented (Table IV) by H = 8 km and 10 km in our working atmospheric model. There is therefore no doubt that in any atmospheric model using diffusion equilibrium at 110 km, the concentrations of all constituents will be correct to a factor of two.

Fig. 7 shows the distribution of the different gases in the thermosphere when diffusion equilibrium has been set up above the level where H = 10 km, i.e. between 110-115 km. The times necessary to achieve such an equilibrium from a state of complete mixing are also shown and it can be seen that more than one day but not more than one week is required.

In the light of these results, it is safe to consider that in the thermosphere above 110 km all minor constituents which are sufficiently inert and neutral to be free from rapid chemical reactions are distributed according to a height gradient corresponding to a diffusion equilibrium. Thus, the density in the thermosphere above a certain altitude depends on the varying molecular mass of the principal constituents. It decreases with height according to the increasing ratio of the concentrations of atomic oxygen and molecular nitrogen, which are the principal constituents.

It may be important to point out that the problem of diffusion of the free charged particles in the ionospheric F2 layer is different from that considered here. The behavior depends not only on the absolute value of the diffusion coefficient[18], but also on several factors[19] such as recombination, ionization and magnetic field for which values must be known as a function of real heights. Atomic oxygen should be an important constituent in the F2 layer, and according to Dalgarno's data[18] for the diffusion coefficient (D) of O^+ in O (about four times the coefficient for neutral particles[19]), diffusion is certainly less marked than previous deductions have indicated.

1.6 Thermal conductivity in the Thermosphere

Convection is the major process of heat transport in the homosphere, particularly in the troposphere, and radiation also becomes important in the stratosphere, but in the thermosphere conduction is of predominant importance. Ionospheric data, rocket results and more recently satellite observations have shown that the temperature gradient must be very large in the thermosphere. It is therefore important to consider the conduction of heat in that part of the atmosphere.

Spitzer[20] has drawn attention to the effect of thermal conductivity in the thermosphere as compared with the heating from absorbed solar radiation. Bates[21] has made an investigation of the thermal equilibrium by considering the rate at which energy is gained from ionizing photons, and lost through conduction and radiation.

Since practically no polyatomic molecule exists in the thermosphere, and since molecular oxygen and nitrogen have no dipole moment, the principal radiation of heat is from atomic oxygen, in virtue of its magnetic dipole emission[21] from the two upper levels of the ground state. Direct application of the heat transfer equation by Bates[22] shows that a large temperature gradient can be maintained only if the heat supplied by photoionization and subsequent energy transformation is sufficiently important. These results do not however lead one to expect a gradient of temperature above the F2 peak because no ultraviolet radiation absorption processes occur there. A discussion by Johnson[23] has shown that if solar electromagnetic radiation is the sole predominant heat source, the temperature cannot increase with height above about 300 km.

Any gradient of temperature above 300 km requires the introduction of heat at higher levels and therefore a transport by conduction throughout the upper thermosphere. Such a source of heating may be supplied by particles entering the atmosphere in various forms. For example, protons of velocities of about 1000 km sec^{-1} having charge transfer cross sections greater than $10^{-16} cm^2$ in nitrogen i.e. between 10 and 50 times the photoionization cross section of ultraviolet radiation, would lead by the ionization process and subsequent dissociative recombination to some heating at heights greater than that by solar radiation(+). However, more evidence on the properties of these particles is needed. Interstellar particles entering the atmosphere could provide a source of heating in the region where their kinetic energy can be transformed into heat by adequate

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(+) $\Delta z \simeq H$, about 4 times the scale height, for a ratio 50 of the cross sections, since $n/n_o = \exp -(z/H)$.

collisions with atmospheric particles(+) .

Other suggestions(++) may be made but Chapman[24] has suggested however that the conduction of heat from the ionized coronal gas to the terrestrial atmosphere could lead to temperature gradients above the ionospheric F_2 peak. In a conservative model of the solar corona he has shown that the heat available is 2.4×10^{19} erg $\sec^{-1}$ at a distance of five earth's radii. If the total energy were trapped, not less than 4 ergs $cm^{-2}sec^{-1}$ would be available at an altitude of 500 km. However, the energy carried down into the thermosphere

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(+) For example, G.A. Harrower, Canadian J. of Phys. 35, 792 (1957) has made an analysis of radio star scintillations in which he considers that the atmosphere would receive about 6×10^{12} hydrogen atoms $cm^{-2}sec^{-1}$ with a velocity of the order of 3×10^6 cm $\sec^{-1}$.

(++) According to Van Allen (to be published, J. Geophys. Res. 1959) there is a possibility that the radiation belt is the source of a leakage of energetic particles contributing to the general heating of the atmosphere.

is only a fraction of the total energy since a concentrative process would be needed. Instead of this, Chapman[24] has indicated that the geomagnetic field would have a hindering effect. It may nevertheless be considered that for the polar regions the magnetic field does not preclude the possibility of the heat transport, nor therefore, the subsequent transport due to the existence of an horizontal temperature gradient.

In conclusion, it appears that sufficient energy may be supplied to the thermosphere from two sources; at the lowest levels up to the F_2 peak by electromagnetic solar radiation and throughout the whole thermosphere by the conduction of heat resulting from far and wide sources such as the coronal gas. Of the processes (convection, radiation and conduction) available for heat loss convection has its major effect in the lower part of the thermosphere, radiation has an effect which decreases with altitude, almost directly in proportion to the atomic oxygen concentration, and conduction remains as the only important process at the highest altitudes. Furthermore, since thermospheric conditions change from the sunlit to the dark atmosphere it must be pointed out that daily variations are unavoidable. Such a variation was discussed by Lowan[25] who made a numerical estimation of the rate of cooling of the thermosphere with a specific temperature gradient after the cessation of the external heating. The calculation showed that the assumed initial height distribution of the temperature begins to shift immediately after the incoming radiation ceases, and the effect is particularly important in the ionospheric F2 layer. In the heat budget of the thermosphere, it is therefore necessary to consider that in addition to a steady state in its upper part, maintained by the heat received by the Chapman's process, there is a daily variation depending on the variation of the incident ultraviolet radiation and on the heat capacity of the thermosphere. In other words, the boundary conditions change between day and night. Since the heating due to solar radiation is caused by processes such as a recombination releasing heat, a night effect cannot be neglected.

If convection is neglected, the heat balance in the thermosphere is a function of the thermal flux arriving at a certain level, the heat production P per unit volume and the heat R radiated per unit volume. Thus, the transfer equation is

$$dE/dz + P = R \qquad (1.60)$$

At the lowest heights in the thermosphere P represents absorption of solar radiation by O_2 molecules undergoing dissociation and transformation of the energy involved in the X-ray spectral region; in the ionospheric F layers, it corresponds to the energy available in the ultraviolet region, for example in the helium lines which according to Oster[26] may lead to more than 10^{10} photons $cm^{-2}sec^{-1}$. Above the peak of the F_2 layer, it is not possible to maintain an active source of heat production.

R symbolizes the energy transformed in the Bates atomic oxygen emission process which is prevalent throughout the thermosphere. It decreases exponentially with height as it follows the vertical distribution of atomic oxygen. The maximum value of R is

$$R = 1.65 \times 10^{-18} n(0) \text{ erg cm}^{-3}\text{sec}^{-1} \tag{1.61}$$

When sufficient energy is supplied by an external heat flow, R is negligible at the highest levels in the thermosphere, and, therefore, the transfer equation is simply

$$dE/dz = 0 \tag{1.62}$$

Conduction related to an energy flux E is expressed by

$$E = \lambda_c \, dT/dz \tag{1.63}$$

where λ_c is the conductivity coefficient.

If we consider a sphere of radius r the total flux F is

$$F = 4\pi f^2 \lambda_c \, dT/dz \tag{1.64}$$

and, instead of (1.62), the transfer equation in the atmosphere must be written

$$dF/dz = 0 \tag{1.65}$$

The coefficient λ_c of heat conduction is connected, following Chapman and Cowling[13], with the coefficient of viscosity μ by an equation

$$\lambda_c = f \mu c_v \tag{1.66}$$

in which c_v denotes the specific heat at constant volume and f is a pure number equal to 2.52 for rigid elastic spheres and 2.50 for Maxwellian molecules. Thus, for a monatomic gas λ_c can be represented with sufficient accuracy by

$$\lambda_c = \frac{5}{2} \mu c_v = \frac{5}{2} \frac{0.1792 (kmT)^{1/2}}{\sigma^2} (3k/2m) \tag{1.67}$$

in which σ is the collision diameter. However, it is customary to represent the temperature variation of viscosity by the empirical expression

$$\mu/\mu_o = (T/T_o)^s \tag{1.68}$$

where s denotes an empirical constant ranging from 0.5 to 1. But, considering that in the thermosphere the range of temperature is different from the range used for the adoption of (1.67), it is easy to see, following Kestin[27], that (1.66) is more precise for thermospheric application. In fact, empirical relations of Keyes[27] show that

$\mu(O_2) = 1.5 \times 10^{-5} T^{1/2}$ within ± 5°/o between 600°K and 1200°K
$\mu(N_2) = 1.25 \times 10^{-5} T^{1/2}$ within ± 5°/o between 600°K and 1500°K
$\mu(\text{air}) = 1.3 \times 10^{-5} T^{1/2}$ within ± 5°/o between 600°K and 1500°K
$\mu(A) = 1.66 \times 10^{-5} T^{1/2}$ within ± 6°/o between 500°K and 1500°K
$\mu(He) = 1.45 \times 10^{-5} T^{1/2}$ within ± 10°/o between 500°K and 1750°K

From these numerical values it may be concluded that (1.66) is applicable in the thermosphere, and thus

$$\lambda_c = A\,T^{1/2} \tag{1.69}$$

giving

$$\lambda_c(\text{air}) = 1.8 \times 10^2 T^{1/2} \tag{1.70}$$

$$\lambda_c(0) = 3.6 \times 10^2 T^{1/2} \tag{1.71}$$

$$\lambda_c(H) = 2.1 \times 10^3 T^{1/2} \tag{1.72}$$

if $\sigma(0) = 2.4$ A and $\sigma(H) = 2.0$ A.

Equation (1.65) can be written using (1.64) and (1.69)

$$4\pi A \frac{d(r^2 T^{1/2} dT/dr)}{dr} = 0 \tag{1.73}$$

and integration between reference levels z and z = 0, corresponding to r = (a + z) and a, the respective distances from the center of the earth, leads to

$$F = 4\pi a(a+z)\,\frac{2}{3}\,A\,\frac{T^{3/2} - T_a{}^{3/2}}{z} \tag{1.74}$$

or

$$E_{a+z} = \frac{a}{a+z}\;\frac{2}{3}\,A\,\frac{T^{3/2} - T_a{}^{3/2}}{z} \tag{1.75}$$

and

$$E_a = E_{a+z}\,(a+z/a)^2 \tag{1.76}$$

If the heat flow is known at a certain distance, a + h , from the center of the earth, (1.74) by using (1.76) leads to

$$\frac{T^{3/2} - T_a{}^{3/2}}{T_h{}^{3/2} - T_a{}^{3/2}} = \frac{z}{h}\,\frac{a+h}{a+z} \tag{1.77}$$

which gives the vertical distribution of the temperature ; (1.77) can be also written as follows

$$T^{3/2} = \frac{(a+h)^2}{a(a+z)} \quad \frac{2\ E_{a+h}}{2A} \ z + T_a^{\ 3/2} \qquad (1.78)$$

which shows that the temperature depends on the boundary condition throughout the term $T_a^{\ 3/2}$.

If x represents a certain distance below or above the height h, (1.78) leads to

$$T_{h+x}^{3/2} + T_{h-x}^{3/2} = x\left[1+(x/a+h)^2\right] 3\ E_{a+h}/A \qquad (1.79)$$

which shows how temperatures are related to the heat flow arriving at a certain specific level.

Expression (1.79) can be applied to a region where atomic oxygen should be the principal constituent i.e. the thermospheric region above the F2 peak. If we consider the heat flow arriving at 500 km and boundary conditions at 300 km, the temperatures at 700 km are as follows :

E_{500km}(erg $cm^{-2}sec^{-1}$)		0.1	0.2	0.5
T_{700km}, if T_{300km}	= 400°K	850	1300	2000
	= 900°K	1250	1550	2300
	= 1600°K	1850	2100	2900

It can be seen that very high temperatures can easily be obtained if a significant fraction of the energy available at the top of the earth's atmosphere is trapped in the thermosphere.

However, the temperature also depends on the lower boundary condition i.e. the conditions below 300 km where absorption of solar radiation is possible. For example, if the temperature is of the order of 1600°K at 300 km, a heat flow of 0.1 erg $cm^{-2}sec^{-1}$ will only lead to a temperature of 1850°K at 700 km i.e. an increase of only 250°K for a range of 400 km, which is not yet detectable.

At such heights, the heat production P per unit volume by ultraviolet radiation is

$$P = n(0)\ K(\lambda)\ E_0(\lambda) \qquad (1.80)$$

in which $K(\lambda)$ denotes the absorption coefficient and $E_0(\lambda)$ the ultraviolet energy available at the top of the earth's atmosphere. Since the absorption coefficient for constituents in the F layer is of the order of 10^{-17} cm^2 the transfer equation (1.60) may be written

$$\frac{dE(\lambda)}{dz} = n(0) \left[10^{-17}\ E_0(\lambda) - 1.65 \times 10^{-18} \right] \qquad (1.81)$$

This shows that the effect of the radiant heat may be neglected when $E_0(\lambda)$ is more than 1 erg $cm^{-2}sec^{-1}$. Because the energy available in the helium lines should be not less than 1 erg $cm^{-2}sec^{-1}$, integration of (1.81) leads to

$$E(\lambda) = E_0(\lambda) \left[1 - e^{-n(0)\ H\ 10^{-17}} \right] \qquad (1.82)$$

which must be compared with the energy supplied by external flow.

A vertical column of less than 10^{16} oxygen atoms cm^{-2} will be easily subject to an external heat flow since, according to (1.82),

$$E(\lambda) < 0.1\ E_0(\lambda) \tag{1.83}$$

It may be concluded, therefore, that the gradient of the scale height in the thermospheric region where only the external heat flow plays a role depends on the conditions prevailing at lower altitudes where the ultraviolet radiation is absorbed.

In the F layer where ultraviolet radiation is absorbed, atomic oxygen and molecular nitrogen are the principal constituents. Using equations (1.70) and (1.71) and taking $g = 900$ cm sec^{-2} equation (1.63) can be written as

$$E = (0.88 \pm 0.06)\ 10^{-3}\ h^{1/2}\ dH/dz \tag{1.84}$$

Fig 8 shows the relationship between heat flow and scale height. For example, a gradient of the order of unity for a scale height of about 50 km corresponds to a heating of 2 erg $cm^{-2} sec^{-1}$ resulting from absorption of ultraviolet radiation above the reference level H = 50 km. Ionospheric observations require that the degree of heating should be higher than this and it is therefore necessary to introduce a constituent which is instrumental in absorption but which does not contribute to the electron concentration as detected by radiotechnics.

1.7. Constitution of the upper atmosphere

The first direct determinations of pressures at altitudes up to 120 km and densities up to 160 km were reported in 1951 by Havens, Koll and LaGow[28] on the basis of about twelve rocket flights. In the same year a conventional atmospheric model combining all available rocket results was adopted by the Rocket Panel[29]. A temperature distribution between 30 and 80 km was deduced which took into account the temperatures observed by all the different groups of observers in the USA. The temperature distribution for the thermosphere up to 220 km was obtained from rocket measurements of the mass densities at about 160 km and 220 km and the temperature gradient adopted for the region between 110 km and 220 km was about 6 deg km^{-1}. The final model was based[30], however, on the assumption of a uniform gradient for the dissociation of oxygen in the 80-120 km range, and for the dissociation of nitrogen in the 120-220 km range. Since the actual height distribution of the oxygen dissociation does not have a uniform gradient and since total dissociation of nitrogen does not occur, the Rocket Panel atmospheric model is not suitable for studying the behavior of the thermosphere.

Table VII shows an atmospheric model for the homosphere between 50 km and 100 km. The temperature at the stratopause is of the order of 273°K, and at the mesopause 190°K or 130°K.

At 100 km, the temperature is 210°K. Comparing the data of Table VII with observational data, it can be said that 273°K is an average value at the stratopause since variations[31] observed there indicate T (stratopause) = 273°K ± 20°K.

Various methods have been used for measuring the density. The results of LaGow, Horowitz and Ainsworth[32] seen to indicate that the arctic densities coincide practically with those at 33°N, i.e. they do not vary by more than 10 to 20 °/ₒ . But according to Jones, Fischbach and Peterson[33], there is a trend for the density at 50 km to decrease with increasing latitude, and for the density at latitude 58°N in the neighborhood of 70 km to vary by a factor of 2, namely between 5×10^{-8} and 10^{-7} gm cm^{-3}.

Observational data for 100 km, are very different(+). The Rocket Panel[30] adopted 4.5×10^{-4} mm Hg, while Havens, Koll and LaGow[28] gave a determination of 4.2×10^{-4} mm Hg. However, Horowitz and LaGow[34] after correcting the pressures measured with Viking 7 have obtained data leading to a value of 1.1×10^{-4} mm Hg, and Byram, Chubb and Friedman[35] indicate from measurements on the incident X-ray flux that the Rocket Panel values were too high by a factor of 3. There is a

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(+) Only pressure measurements were made. There is no direct measurement of density.

variation of a factor of 4 for all these values, which are for White Sands (33°N). The data (USSR) published by Mikhnevitch[36] indicate a relatively low pressure : 1.8 x 10^{-4} mm Hg at 100 km. But recent data obtained at Fort Churchill (58°N) show that the pressure[37] is not less than 3 x 10^{-4} mm Hg and that density data[38] for molecular oxygen must lead to a total density of not less than 9 x 10^{-10} gm cm^{-3}. Thus, there is a broad range of a factor four in the pressure and density data at 100 km. This cannot be attributed to a permanent latitudinal effect since the variation at White Sands corresponds to the difference between the White Sands minimum value and the average values obtained at Fort Churchill. In other words, the pressure at 100 km can be of the order of (3 ± 1) x 10^{-4} mm Hg, but it is not yet possible to determine the exact range of real variations.

Since any atmospheric model, even at such low heights as 100 km, is subject to uncertainty, we have introduced (Table VII) between 70 km and 100 km the variation of temperature which is required to lead to a low pressure at 100 km. Because large variations of the temperature may occur near the mesopause, the layer between 70 km and 100 km is divided in two parts, 15 km thick, in order to obtain the minimum temperature near 85 km. In any case, with a pressure of 4.5 x 10^{-2} mm Hg corresponding to a density of 10^{-7} gm cm^{-3} at 70 km (see

Table VII), and if the pressure at 100 km varies by a factor of about 3 there would be a variation of about 60°K at 85 km. If the temperature at 70 km and 100 km is 210°K, the general trend should be as follows :

Temperature at 85 km (°K)

130	140	150	160	170	180	190	200	210

Pressure at 100 km (10^{-4} mm Hg)

1.1	1.4	1.7	2.0	2.3	2.6	3.0	3.4	3.8

If there is a possibility of a very large variation of the pressure at 100 km, any atmospheric model to be used for the thermosphere above 100 km will be affected to the same extent. For example, a density at 200 km of the order of 6×10^{-13} gm cm^{-3} (deduced from satellite observations) requires a completely different structure of the atmosphere between 100 km and 200 km if the density at 100 km varies by a factor of 3.

The data available on atmospheric density, obtained by means of rockets and of satellites, must be compared even if sufficient data is still lacking. Present rocket information[37,39] on densities at 200 km (59°N) indicates a variation by a factor greater than 10 between a summer day (6×10^{-13} gm cm^{-3}) and a winter night (4×10^{-14} gm cm^{-3}). Analysis of satellite observations made by several authors[40-55] shows that in general the density at such altitudes does not vary by more than a factor 2.

Fig. 9 shows the variations in the density[46] at the altitude of the perigee of Sputnik II. The period of revolution may be affected by complex phenomena, such as the orientation of the object and particular drag coefficients. Irregularities due to these effects, however, do not seem to be the most important, for they can be explained by variations in the atmospheric density at the altitude of the perigee. Furthermore, these irregularities in the decceleration (Fig. 9), which are within ± 20 °/₀ of the average value, show that in the range of altitudes covered by the perigee of Sputnik II (initial perigee, 240 km) the actual value of the density is always equivalent to the average value ± 20 °/₀. A certain variation in the indicated limits may result from a difference due to the position of the perigee in the night-side or the day-side of the earth, but it is clear that the most obvious association must be made with the variation of the solar radiation which causes the heating of the thermosphere.

Fig. 10 shows the variation of density[48] at the altitudes of the perigees of the Vanguard satellite and Sputnik III. The simulta-simultaneous variations in the densities of the atmosphere near 650 km (± 50 °/₀) and 250 km (± 20 °/₀) must result from a change in the atmospheric conditions below 150 km to which the heat is conducted from higher levels where the ultra-violet radiation is absorbed, and perhaps in addition from a variation in the heat flow coming from the top of the atmosphere.

Analysis of satellite observations such as those made by Sterne[40], Harris and Jastrow[45], Siry[55], Jacchia[47], Schilling and Witney[49] shows that densities between 200 km and 800 km may have the following vertical distribution : (see Fig.11),

Altitude (km)	Density (gm cm^{-3})	Density (gm cm^{-3})
200	6×10^{-13}	
250		2×10^{-13}
300	6×10^{-14}	
350		2×10^{-14}
420	6×10^{-15}	
500		2×10^{-15}
600	6×10^{-16}	
700		2×10^{-16}
800	6×10^{-17}	

Such a vertical distribution of thermospheric densities must be interpreted by an increase of the atmospheric scale height between 200 km and 800 km. According to formula (1.6)

$$d\rho / \rho = - (1+1/\beta)\, dH/H \qquad (1.85)$$

in which

$$\beta = dH/dz \tag{1.86}$$

and

$$dH/H = dT/T - dm/m - dg/g \tag{1.87}$$

If the variation of the scale height were due only to the variation of temperature(+), the gradient β should be given, according to (1.80), by

$$\beta \simeq 10^3 \; E/H^{1/2} \tag{1.88}$$

The heat flow E depends on the absorption processes of the ultraviolet radiation below 300 km and on the transport by conduction in the upper part of the thermosphere.

If the variation of the scale height were due only to the variation of the mean molecular mass(++) m, the gradient β should be

$$\beta \simeq -(1/m^2) \; dm/dz \tag{1.89}$$

depending on the level at which diffusion begins.

Thus, when a complete interpretation of the thermospheric densities is required, it is necessary to consider the simultaneous variations of 4 parameters z, T, m and g in the

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(+) This is the case when, if the variation of g is neglected, no diffusion is considered : m = constant.

(++) This is the case when, the variation of g being neglected, diffusion plays a role in an isothermal atmosphere.

following expression

$$d\rho/\rho = -\,dz/H - dT/T + dm/m + dg/g \qquad (1.90)$$

It is clear that an exact knowledge of the composition and constitution of the thermosphere cannot be obtained without knowing the simultaneous variations of T and m. In other words, there are many possibilities for thermospheric models leading to the densities obtained from satellite measurements. Among these models a few can be selected by lower boundary conditions, namely by assumptions made on temperature and mean molecular mass between 100 km and 200 km.

In order to show how different interpretations can be obtained, we shall show how various conditions can be introduced between 100 and 200 km and then adopt two extreme models.

In Table VII two pressures, 1.1×10^{-4} and 3×10^{-4} mm Hg, are obtained at 100 km according to conditions required at the mesopause. These pressures were computed for the homosphere in which no dissociation of molecular oxygen occurs. Making the following assumptions :

Model A	Model B
Strong dissociation of O_2:80°/o	Low dissociation of O_2:25°/o
$n(O) \quad n(N_2)$	$n(N_2) + n(O_2) \quad n(O)$,

the conditions at 100 km are

$p = 1.4 \times 10^{-4}$ mm Hg	$p = 3 \times 10^{-4}$ mm Hg
$T = 200°K$	$T = 200°K$
$\rho = 2.9 \times 10^{-3}$ gm cm^{-3}	$\rho = 6.6 \times 10^{-10}$ gm cm^{-3}
$M = 25.5$	$M = 27.4$

Between 100 km and 150 km various conditions can be adopted. We choose the following ones :

Model A			Model B		
Diffusion from 110 km,			No diffusion,		
Temperature at 150 km			Temperature at 150 km		
(a)	(b)	(c)	(a)	(b)	(c)
450°K	575°K	700°K	1180°K	1650°K	2100°K
leading to					
$\rho = 2.0 \times 10^{-12}$	2.9×10^{-12}	3.2×10^{-12}	$\rho = 1.55 \times 10^{-12}$	1.5×10^{-12}	1.4×10^{-12}
$M = 18.0$	18.9	19.2	$M = 27.4$	27.4	27.4

These two extreme cases show how the density depends on the mean molecular mass and temperature. It is clear however that the variation of density is not very sensitive to the variation of temperature in the lower thermosphere since $\rho = (2.3 \pm 0.9) \times 10^{-12}$ gm cm^{-3} in all circumstances. As an example, Fig. 12 exhibits the density-altitude relation for Model B above 120 km for three different large gradients of the scale height, namely $\beta = 1$, 1.5 and 2.0 corresponding at

150 km to heat flows of the order of 1.7, 2.0 and 2.5 erg cm^{-2} sec^{-1}, respectively. It is found by the procedure adopted that solutions can be obtained which are close to observed values but which relate to an aeronomic problem impossible to resolve. The procedure leads however at higher altitudes to different solutions such as the following ones at 200 km :

	Model A			Model B		
	(a)	(b)	(c)	(a)	(b)	(c)
ρ =	$2.1\mathrm{x}10^{-13}$	$4.3\mathrm{x}10^{-13}$	$6.1\mathrm{x}10^{-13}$	$3.8\mathrm{x}10^{-13}$	$5.2\mathrm{x}10^{-13}$	$6.1\mathrm{x}10^{-13}$

Here one can exclude one or another possibility if the atmospheric density at 200 km does not vary between 2×10^{-13} and 6×10^{-13} gm cm^{-3}.

For the sake of simplicity we will now discuss each model separately, since Model A requires a heat flow conducted from the top of the earth's atmosphere and Model B is based on the assumption of heating due only to photo-ionization and subsequentdissociative recombination.

Fig. 13 shows the variation of density in the thermosphere up to the 800 km level, the heat flow being between 0.1 and 1.0 erg $cm^{-2}sec^{-1}$. The basic boundary conditions are taken to be T = 700°K at 150 km and the particle concentrations are deduced from Model A at 100 km where $n(O) \simeq n(N_2)$. Attention must be drawn to the portion of Fig. 13 which refers to the region between 200 km and 300 km since it shows that the density is not very sensitive to variations of the heat flow. But great significance must be attached to the behavior of the atmosphere below 150 km, i.e. to boundary conditions. In the region above 300 km the variation of the density is more pronounced and depends primarily on the importance of the heat flow. Data from satellites applied to such a working model suggest a heat flow of the order of 0.15 erg $cm^{-2}sec^{-1}$. A variation of a factor of 5 in the density at 650 km corresponds to fluctuations between 0.15 and 0.3 erg $cm^{-2}sec^{-1}$.

Returning to Model B, it can be shown that temperatures of the order of 1650°K and 2100°K at 150 km lead to densities at 600 km which are too high, namely more than 10^{-15} gm cm^{-3}. Attention is therefore directed to the analysis of the model for which the temperature is not more than 1200°K at 150 km.

Fig. 14 shows the variation of density with diffusion beginning at 110 km or 150 km, when the temperature is constant above 150 km or 220 km respectively. In the atmospheric region covered by satellite observations the fit is close over the major part of the curve if a constant temperature is assumed. The lower part, below 250 km, is, however, not adequately represented. Nevertheless since reliable information on variations is lacking, differences of 50 °/o may be accepted. Fig. 15 exhibits the relations between scale height and mean molecular mass between 100 km and 700 km in order to show how the basic parameters vary with height. In the region below 150 km the variation of H is related to the variation of T, while above 220 km the two parameters H and m are closely related. Since reliable information on the variation of the scale height with altitude is lacking, several solutions for the vertical distribution of the density are possible, depending on conditions chosen below 200 km.

The discussion of the two atmospheric models concerns physical conditions which are completely different, and the observed values of the density in the thermosphere can be approached without difficulty using either assumption. Any attempt at distinguishing an effect of heating due only to ultraviolet radiation, i.e. T = constant at high altitudes, and the addition of an external heat flow, i.e. T increasing with height, meets with several difficulties. From energies which are involved in both cases it is however apparent that the total heating rate near the F_1 peak is strongly dependent on the heating due to solar radiation. This leads to a temperature gradient which must be greater than 20 deg km^{-1}.

If the number of photons available in the resonance line of He II, at $\lambda = 304$ A, is of the order of 10^{11} photons cm^{-2} sec^{-1} i.e. an energy of about 4 ergs $cm^{-2} sec^{-1}$ available for heating, the rate of ionization γb at the absorption peak is according to Nicolet's formula[14]

$$\mathcal{H} = (1+\beta)\, e^{-(1+\beta)}\, Q_\infty / H \qquad (1.91)$$

in which Q_∞ is the number of photons available at the top of the earth's atmosphere. A local scale height $H \gtrsim 50$ km and a gradient $\beta = 1.5$ leads to a production rate of electrons at the F_1 peak of the order of

$$\mathcal{H} = 3000 \pm 1000 \text{ electrons cm}^{-3}\text{sec}^{-1} \qquad (1.92)$$

which is not less than 10 times the production rate deduced from ionospheric observations. The conventional method used in the analysis of ionospheric data gives

$$q = \alpha\, n_e^2 \simeq 300 \text{ electrons cm}^{-3}\text{sec}^{-1} \qquad (1.93)$$

at the peak of the F_1 layer.

As has been pointed out before[22], it is necessary that at least two species of ions are formed and recombine with different recombination coefficients. Here, molecular nitrogen and atomic oxygen are involved. Since the time of recombination of $\tau_{rec}(N_2^+)$ is certainly less (100 to 1000 times) than that of $\tau_{rec}(O^+)$, the following conditions must be adopted near the F_1 peak

$$n(N_2) \geqslant 10\, n(O) \qquad (1.94)$$

in order to reconcilie (1.92) with (1.93). Hence one can exclude the possibility that atomic oxygen(+) is the principal constituent of the thermosphere below 200 km, for according to the above the percentage of dissociation of molecular oxygen should be less than 20 °/₀. In other words, if condition (1.94) must be applied, the fraction of dissociated oxygen is small, and at 150 km

$$n(O_2) \gg 2n(O) \qquad (1.95)$$

The discussion in the two preceding paragraphs concerns steady conditions in the thermosphere. The variable heating during a solar cycle will not, however, change the conditions if the atmosphere in the E layer is not affected (see Fig. 12.). Much more important effects may be expected from variations between day and night. Since the heat transfer is governed by the differential equation

$$\frac{\partial T}{\partial t} = \frac{5}{2nm} \operatorname{div} \left[\mu \left(\operatorname{grad} T\right) \right] \qquad (1.96)$$

or, in one-dimensional conduction transport (1.67 and 1.69),

(+) In the F_2 layer the production of electrons due to atomic oxygen should increase compared with that resulting from molecular nitrogen, for the ratio $n(O)/n(N_2)$ increases with height. Such an effect should be introduced in the analysis of the behavior of the F_1 and F_2 layers.

$$\frac{\partial T}{\partial t} = \frac{\overline{A\ T}^{1/2}}{(5/2)k\ n} \quad \frac{\partial^2 T}{\partial x^2} \tag{1.97}$$

it is necessary to determine the periods of time involved in the heat transport as a function of the concentrations and distances.

As an example, let us consider an atomic oxygen atmosphere with two regions in which the temperatures, T_0 and T_1 are such that $T_0 = 2T_1$. The time, $\tau_{\text{conduction}}$, necessary to increase T_1 to $T = 0.9\ T_0$ can be deduced from (1.97). The solution(+) of (1.97) is

$$\tau_{\text{conduction}} = 2 \times 10^{-19} x^2 n \tag{1.98}$$

According to (1.98) a temperature equilibrium to a distance of 10,000 km would be reached in 2×10^4 sec with a concentration, n, of the order of 10^5cm^{-3}. This shows that it is not possible to maintain a difference of temperature with latitude at very high altitude. At altitudes below 200 km, however, where $n > 10^{10}$ cm^{-3} a temperature equilibrium would only be obtained to a maximum distance of 30 km in the same time (2×10^4 sec.).

It is therefore clear from the times involved in the conduction of heat in the thermosphere that densities at very high altitudes must depend primarily on the temperature near 200 km. Since it has been shown that the density at 200 km is not very sensitive to the gradient of temperature, the variations at such

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(+) The solution would be the same if $T_1 = 0.9\ T_0$ and $T = 0.98\ T_0$ or $T_1 = 0.8\ T_0$ and $T = 0.96\ T_0$

an altitude should be relatively small, but the magnitude of the variations should increase with altitude. The analysis of satellite observations made by Jacchia shows that the oscillations of the solar radiations are amplified from 200 km to 700 km, and therefore demonstrates the direct effect of ultraviolet radiation in heating the thermosphere. On the other hand, the importance of these variations compared with the variations resulting from a possible latitude effect leads us to the conclusion that there is a strong tendency to isothermy with latitude at the highest altitudes. Similarly, a variation between day and night will be related to this tendency to isothermy at the highest latitudes but should depend on the temperature near 200 km, because the density above that altitude depends essentially on the absolute value of the temperature.

In conclusion, valuable information on the vertical structure of the thermosphere would be obtained from a complete study of the whole region above 100 km. Additional data from rockets for the region between 100 km and 200 km are essential to know the exact composition. Between 200 and 1000 km, continuous observations from the variation of the acceleration of satellites ($\alpha \rho H^{1/2}$) would indicate how $\rho H^{1/2}$ varies with solar radiation in sunlit and dark atmospheres. Using these methods, an important parameter of the thermosphere, i.e. the temperature, will be determined, with its daily variation and its vertical distribution. Such a determination is essential to avoid explanations of the composition and constitution of the thermosphere based only on working atmospheric models.

TABLE 1. Principal Constituents of the Air at Ground Level

Molecule	Mass(1)	Percentage	Concentration(2)	Total(3)
N_2	28.022	78.084 ± 0.004	2.098×10^{19}	1.678×10^{25}
O_2	32.009	20.946 ± 0.002	5.629×10^{18}	4.501×10^{24}
A	39.960	0.934	2.510×10^{17}	2.007×10^{23}
CO_2	44.024	0.33	8.87×10^{15}	7.09×10^{21}
Air	28.973	100.00	2.687×10^{19}	2.148×10^{25}

(1) Physical Mass with Avogadro's number : 6.025×10^{23} gm mol^{-1}.

(2) Number of molecules cm^{-3} at standard pressure (760 mm Hg) and standard temperature (273.16°K). Loschmidt's number : $2.687 \times 10^{19} cm^{-3}$.

(3) Number of molecules per cm^2 column, with a reduced height or scale height defined by $H = kT/mg$ where $k = 1.38 \times 10^{-16}$ erg deg^{-1}, $T = 273.16°K$, $m = 48.08 \times 10^{-24}$ and $g = 980.665$ cm sec^{-2}.

TABLE II. Inert Gases at Ground Level

Atoms	Ratio by Volume	Number (cm^{-3})	Number (cm^{-2})
He	$5.24 \pm 0.004 \times 10^{-6}$	1.41×10^{14}	1.13×10^{20}
Ne	$1.818 \pm 0.004 \times 10^{-5}$	4.89×10^{14}	3.89×10^{20}
A	$9.34 \pm 0.01 \times 10^{-3}$	2.51×10^{17}	2.01×10^{23}
Kr	$1.14 \pm 0.01 \times 10^{-6}$	3.06×10^{13}	2.45×10^{19}
Xe	$8.7 \pm 0.1 \times 10^{-8}$	2.34×10^{12}	1.87×10^{18}

TABLE III. Molecular Content of Minor Constituents

Molecule	Ratio by Volume	Remarks
H_2O	10^{-5} to 10^{-2}	Variable, particularly in troposphere. Decrease in stratosphere. Dissociation in mesosphere.
O_3	10^{-7} to 10^{-8}	Increase with height, maximum in stratosphere and diurnal variation in mesosphere.
CH_4	1.5×10^{-6}	Mixed in troposphere, oxydized in stratosphere. Dissociation in mesosphere.
N_2O	5×10^{-7}	Mixed in troposphere. Continuous dissociation in stratosphere and mesosphere.
CO	6×10^{-8} to 2×10^{-7}	Variable.
H_2	5×10^{-7}	Mixed. Dissociation in mesosphere.
NO_2 NO	5×10^{-10} to 2×10^{-8}	Industrial origin in troposphere. Photochemical origin in mesosphere and thermosphere.

TABLE IV. Oxygen Dissociation Rate Coefficients* (in sec^{-1})

Black Body T / Molecules cm^{-2}	4000°K	4500°K	5000°K	5500°K	6000°K
10^{15}	1.24×10^{-7}	1.57×10^{-6}	1.28×10^{-5}	6.64×10^{-5}	2.76×10^{-4}
10^{16}	1.18×10^{-7}	1.47×10^{-6}	1.15×10^{-5}	6.19×10^{-5}	2.56×10^{-4}
10^{17}	7.34×10^{-8}	8.63×10^{-7}	6.43×10^{-6}	3.34×10^{-5}	1.34×10^{-4}

(*) This table shows the rates at which O_2 would be dissociated if it were irradiated with black body radiation having the different temperatures listed and subject to preliminary absorption by columns of 10^{15}, 10^{16} and 10^{17} O_2 molecules cm^{-2}.

TABLE V. Life Times of Oyxgen Atoms

n (M)$^{(*)}$ n(O)	10^{15}	10^{14}	10^{13}	10^{12}(cm^{-3})
10^{13}(cm^{-3})	<1 day	<1 day	1 week	
10^{12}	<1 day	1 week	>1 month	>1 year
10^{11}	1 week	>1 month	>1 year	>10 years
10^{10}	>1 month	>1 year	>10 years	

(*) This table shows how the time for the formation of molecules from oxygen atoms, by a recombination process involving a third particle (M), depends on the concentration of M and of the original concentration of O.

TABLE VI. Working Atmospheric Model for the Study of Diffusion

Scale Height (km)	Altitude(+) (km)	Temperature (°K)	Pressure (mm Hg)	Concentration (cm^{-3})
8	100 - 105	220	1.1×10^{-4}	4.8×10^{12}
10	110 - 115	272	3.5×10^{-5}	1.25×10^{12}
12	120 - 125	326	2.2×10^{-5}	4.2×10^{11}
14	130 - 135	379	9.4×10^{-6}	1.7×10^{11}
16	140 - 145	432	3.3×10^{-6}	7.5×10^{10}

(+) This table corresponds to an atmospheric model in which the gradient of the scale height $dH/dz = 0.2$ and the mean molecular mass $M = 24$, i.e. $n(N_2) = 2\, n(O)$. Observations show that a pressure of the order of 10^{-4} mm Hg corresponds to altitudes between 100 km and 105 km.

TABLE VII. Atmospheric Data between 50 and 100 km.

Altitude (km)	Temperature[(+)] (°K)	Pressure (mm Hg)	Concentration (cm^{-3})	Density ($gm\ cm^{-3}$)
50.0	274	6.7×10^{-1}	2.4×10^{16}	1.1×10^{-6}
52.5	274	4.9×10^{-1}	1.7×10^{16}	8.3×10^{-7}
55.0	274	3.6×10^{-1}	1.3×10^{16}	6.1×10^{-7}
57.5	263	2.7×10^{-1}	9.7×10^{15}	4.7×10^{-7}
60.0	253	1.9×10^{-1}	7.3×10^{15}	3.5×10^{-7}
62.5	242	1.4×10^{-1}	5.4×10^{15}	2.6×10^{-7}
65.0	232	9.6×10^{-2}	4.0×10^{15}	1.9×10^{-7}
67.5	221	6.6×10^{-2}	2.9×10^{15}	1.4×10^{-7}
70.0	210	4.5×10^{-2}	2.1×10^{15}	9.9×10^{-8}
72.5	197	3.0×10^{-2}		7.0×10^{-8}
	207	3.0×10^{-2}	1.4×10^{15}	6.7×10^{-8}
75.0	183	1.9×10^{-2}		4.9×10^{-8}
	203	2.0×10^{-2}	9.5×10^{14}	4.6×10^{-8}
77.5	170	1.2×10^{-2}		3.3×10^{-8}
	200	1.3×10^{-2}	6.4×10^{14}	3.1×10^{-8}
80.0	156	7.2×10^{-3}		2.1×10^{-8}
	197	8.7×10^{-3}	4.3×10^{14}	2.1×10^{-8}

(+) Between 70 km and 100 km, two distributions of temperature for atmospheres in which the temperatures at 85 km are 130°K and 190°K.

TABLE VII. Atmospheric Data between 50 and 100 km. (Contd)

Altitude (km)	Temperature (°K)	Pressure (mm Hg)	Concentration (cm^{-3})	Density (gm cm^{-3})
82.5	143	4.1×10^{-3}		1.3×10^{-8}
	193	5.7×10^{-3}	2.8×10^{14}	1.4×10^{-8}
85.0	130	2.2×10^{-3}		8.0×10^{-9}
	190	3.7×10^{-3}	1.9×10^{14}	9.0×10^{-9}
87.5	143	1.2×10^{-3}		3.9×10^{-9}
	193	2.4×10^{-3}	1.2×10^{14}	5.7×10^{-9}
90.0	156	7.0×10^{-4}		2.1×10^{-9}
	197	1.6×10^{-3}	7.6×10^{13}	3.7×10^{-9}
92.5	170	4.2×10^{-4}		1.1×10^{-9}
	200	1.0×10^{-3}	4.9×10^{13}	2.4×10^{-9}
95.0	183	2.6×10^{-4}		6.6×10^{-10}
	203	6.8×10^{-4}	3.2×10^{13}	1.5×10^{-9}
97.5	197	1.7×10^{-4}		4.0×10^{-10}
	207	4.5×10^{-4}	2.1×10^{13}	1.0×10^{-9}
100.0	210	1.1×10^{-4}		2.5×10^{-10}
	210	3.0×10^{-4}	1.4×10^{13}	6.6×10^{-10}

Legends of Figures

Fig. 1. The main names of the atmospheric regions, from troposphere to thermosphere, are based on a thermal classification. When the classification is given according to composition, the term homosphere means that the mean molecular mass is constant, and heterosphere means that the mean molecular mass varies. Names such as ionosphere, chemosphere, ozonosphere, ... , are descriptive, of regions where particular processes take place.

Fig. 2. Possible vertical distributions of atomic and molecular oxygen in the lower thermosphere.

Because the times involved in the recombination of oxygen increase rapidly with height above the mesopause, photoequilibrium cannot be maintained. The altitude of the peak of atomic oxygen is not determined by photochemical equilibrium conditions but by the effect of a downward transport depending on mixing conditions. The figure shows how the conditions at the atomic oxygen peak vary from photo-equilibrium to strong mixing. It must be pointed out that in this figure the altitude of 100 km corresponds to a pressure of 10^{-4} mm Hg. Such a pressure may correspond to 105 km in the atmosphere described in Tables VI and VII, and the concentrations indicated in the figure between 90 km and 110 km can therefore represent conditions between 95 km and 115 km when the pressure at 100 km is 3×10^{-4} mm Hg.

Fig. 3. Photo-equilibrium of molecular oxygen changed into diffusive equilibrium by upward transport of O_2. Height is represented by the scale height appropriate to each level in the atmosphere described in Table VI. The abscissae represent the factor by which the initial concentration $n(O_2)_{t=0}$, appropriate to photo-equilibrium alone, is increased to give a concentration $n(O_2)_t$, after a time t, which is different for each of the curves. The points O, □ , X and + represent characteristic distributions of O_2 in diffusive equilibrium for 4 different boundary conditions. For example, if diffusion begins at H = 10 km (symbol O and corresponding to altitudes between 100 and 105 km), a diffusion distribution of O_2 will be reached above this reference level in not more than two days.

Fig. 4. Mixing distribution of molecular oxygen changed into diffusive equilibrium by downward transport of O_2. Height is represented by the scale height appropriate to each level in the atmosphere described in Table VI. The abscissae represent the factor by which the initial concentration $n(O_2)_{t=0}$, appropriate to mixing alone, is <u>decreased</u> to give the final concentration $n(O_2)_t$ after a time t, which is different for each of the curves. The points Δ, O, and □ represent characteristic distributions of O_2 in diffusive equilibrium for 3 different boundary conditions. For example, if diffusion begins at H = 10 km (symbol O) and is prevented

below this reference level, molecular oxygen will reach a vertical diffusive distribution above the reference level in not more than three days.

Fig. 5. Diffusion of argon in the thermosphere.

Mixing distribution of argon changed into diffusive equilibrium by downward transport. Height is represented by the scale height appropriate to each level in the atmosphere described in Table VI. The abscissae represent the factor by which the initial concentration $n(A)_{t=0}$, appropriate to mixing alone, is decreased to give the final concentration $n(A)_t$ after a time t, which is different for each of the curves. The points △ , 0 and □ represent characteristic distributions of argon in diffusive equilibrium for 3 different boundary conditions, namely H = 12.5 km, 10 km and 8 km. These distributions indicate that diffusive equilibrium distributions of argon can be obtained above reference levels H = 12.5 km, 10 km and 8 km in not more than 1, 3 and 7 days, respectively. Since the reference level H = 10 km corresponds to altitudes in the range 100-105 km in which observational data show diffusive separation of A and N_2, the time of mixing is more than 3 days, and less than 7 days at H = 10 km.

Fig. 6. Diffusion of helium in the thermosphere.

Mixing distribution of helium changed into diffusive equilibrium by upward transport. The abscissae represent the factor by which the initial concentration $n(He)_{t=0}$, appropriate to mixing alone, is increased to give the final concentration $n(He)_t$ after a time t, which is different for each of the curves. The symbols have the same meaning as those of Fig. 5. Comparing results of Fig. 5 and 6, it can be seen that the times required for diffusion of helium are about twice those for the diffusion of argon.

Fig. 7. Change of mixing distribution of various minor constituents into diffusive distribution by upward or downward transport.

Height is represented by the scale height appropriate to each level in the atmosphere as described in Table VI. H = 10 km is the reference level corresponding to altitudes between 100 and 105 km. Below the reference level mixing prevents any separation of the constituents while diffusion exists above this level. Such conditions are obtained in less than 3 days for all minor constituents (symbol O) except for helium which requires 7 days (symbol △).

Fig. 8. Heat Flow and Scale Height.

The effect of thermal conductivity in the thermosphere is shown by a curve relating the scale height and the energy necessary to maintain the gradient $\beta = dH/dz$. From the abscissae representing E/β, it may be deduced that any gradient β of the order of unity requires energy flows from 1 to 3 ergs $cm^{-2}sec^{-1}$

in the whole thermosphere since thermospheric scale heights are between 10 km and 100 km. It must be pointed out that the distribution of E with height must follow a distribution given by an equation of a type described by (1.82).

Fig. 9. Orbital acceleration in seconds per day of Satellite 1957 β1 (Sputnik II) during five months, according to Jacchia[46]. The perigee, which was about 225 km during the first revolution, changed slowly from 50°N to the equator. The fluctuations of the acceleration around an average value, which are less than 20 ‰, may be explained by the variations in the solar heating, reaching a maximum near 150 km. A first analysis may be made in considering the position of the perigee in a sunlit or dark atmosphere. However, according to Jacchia (Harvard College Observatory, Announcement Card 1423, January 16, 1959) a comparison can be made between the solar radiation represented by its emission at 2800 Mc/s and the variation of the orbital acceleration of the satellites. Solar radiation data are taken from the Bulletin of the National Research Council, Ottawa, Canada.

Fig. 10a. Orbital acceleration in days per day of Satellite 1958 1 (Sputnik III) during five months, according to Jacchia (private communication). During 2500 revolutions the orbital acceleration is subject to rhythmic variations of less than

20 °/$_o$. There is evidence that these fluctuations are associated with the principal variations of the solar radiation represented by the solar flux at 2800 Mc/s. The associations between variations of the atmospheric density and the solar heating is more apparent in Fig. 10b.

Fig. 10b. Comparison between the variations of the orbital acceleration of satellites Vanguard and Sputnik III in association with the solar heating, according to Jacchia's data. During 1500 revolutions (about 5 months) the fluctuations of the acceleration of the Satellite 1958 β2 (Vanguard) lead, near the perigee, at an altitude of 656 km, to variations of about 50 °/$_o$ in the density, namely $\rho = (3.6 \pm 1.8) \times 10^{-16}$ gm cm^{-3}.

A sharp increase in the acceleration in August corresponds to an increase of a factor 2.5 in the average density between the two periods April - July and September - December. However, the fluctuations are related to variations of the solar radiation. As an example the principal peak in October is well represented in the three curves, indicating that the accelerations of both satellites varied almost in unison with the solar heating.

Fig. 11. Density-altitude relations deduced from satellite observations.

The observational values are taken from Sterne[40] for Sputnik II (1957 α2) and Explorer I (1958 α), from Schilling and Whitney[49] for Explorer IV (1958 ε), from Harris and Jastrow[45] for Explorer I and Vanguard (1958 β2), Jacchia[47] for Vanguard and Siry[55] for Explorer I and Vanguard.

Fig. 12. Density-altitude relations above 120 km for various heat flows at 150 km and constant scale height gradients. The symbols (X) represent the data from a rocket flight at Fort Churchill (59°N) published by LaGow, Horowitz and Ainsworth[37]. The height variation of the density shows that any solution for a too short range of altitudes is highly arbitrary.

Fig. 13. Effect of heat flows, conducted from the top of the earth's atmosphere, on thermospheric densities.

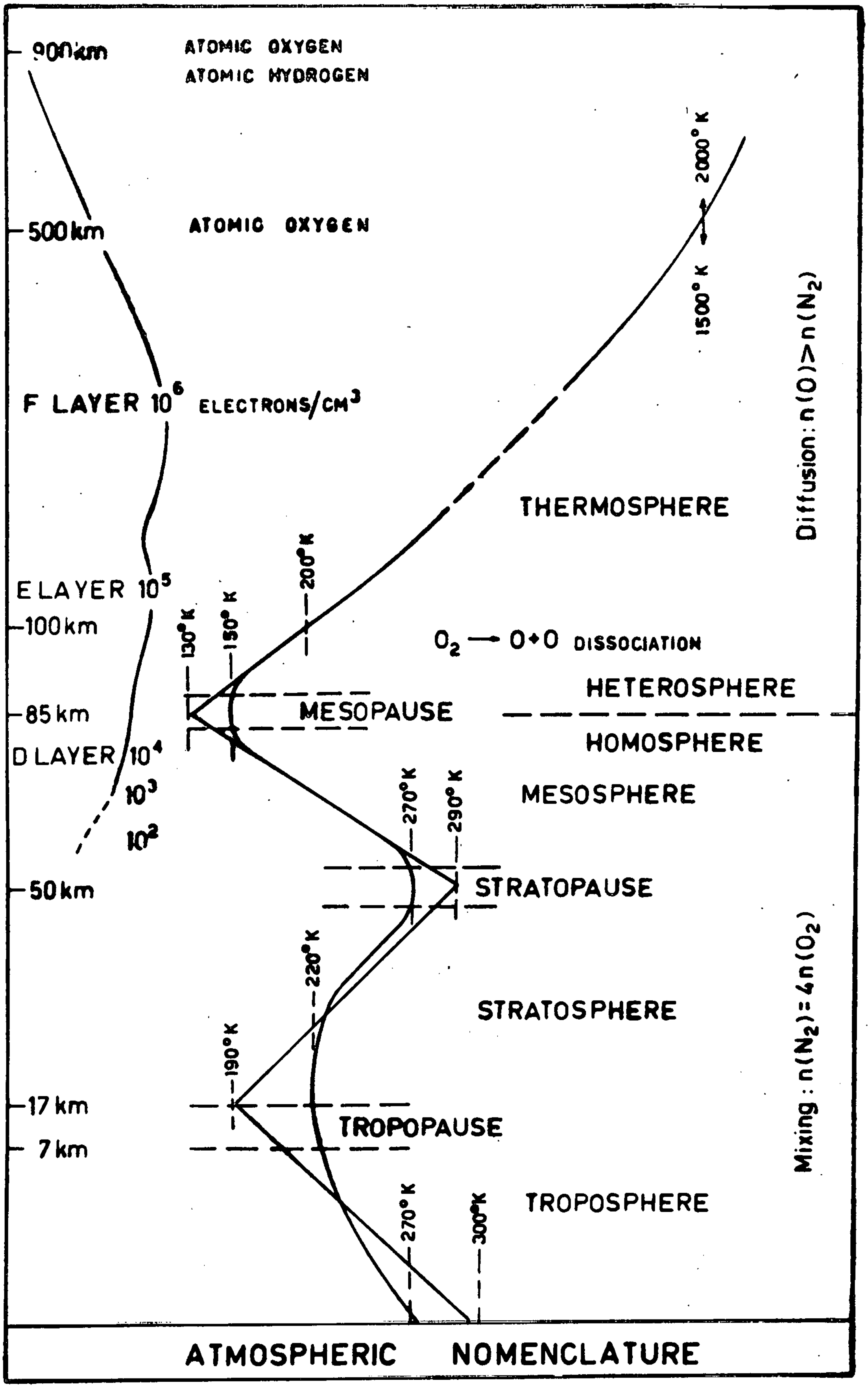
If one adopts at 500 km a rate of conduction of energy in the range from 0.1 to 1.0 erg $cm^{-2}sec^{-1}$ it can be seen that the thermospheric densities are very sensitive to the energy input. The absolute values of density depend on the lower boundary condition at 150 km where $T = 700°K$ and $n(O) > n(N_2)$. The crosses indicate the satellite data. Between 200 km and 300 km the density is not very sensitive to the energies which are involved in the heat flow. Variations observed at 650 km would correspond to a variation of a factor 2 in the heat flow. This figure corresponds to Model A discussed in the text.

Fig. 14. Densities in an isothermal atmosphere.

The observational curve can be followed when certain conditions are assumed below 200 km. Assuming large gradients of the temperature between 120 km and 160 km (see Fig. 12) the effect of diffusion will be to lead to a decrease with altitude of the mean molecular mass corresponding to an increase of the scale height. It is possible to fit the satellite observations of density above 200 km to this curve. This figure corresponds to Model B discussed in the text.

Fig. 15. Variations of the scale height and of the mean molecular mass with altitude.

When no gradient of the temperature is assumed above a certain altitude (here 220 km), the vertical distribution of density deduced from satellite observations (see Fig 11) can be explained only if the scale height increases with height. Thus, the variation of the scale height must be associated with a variation of the mean molecular mass depending on diffusion. The relation scale height - mean molecular mass of this figure must be associated with the density shown in Fig. 14.



ATMOSPHERIC NOMENCLATURE

Figure 1.

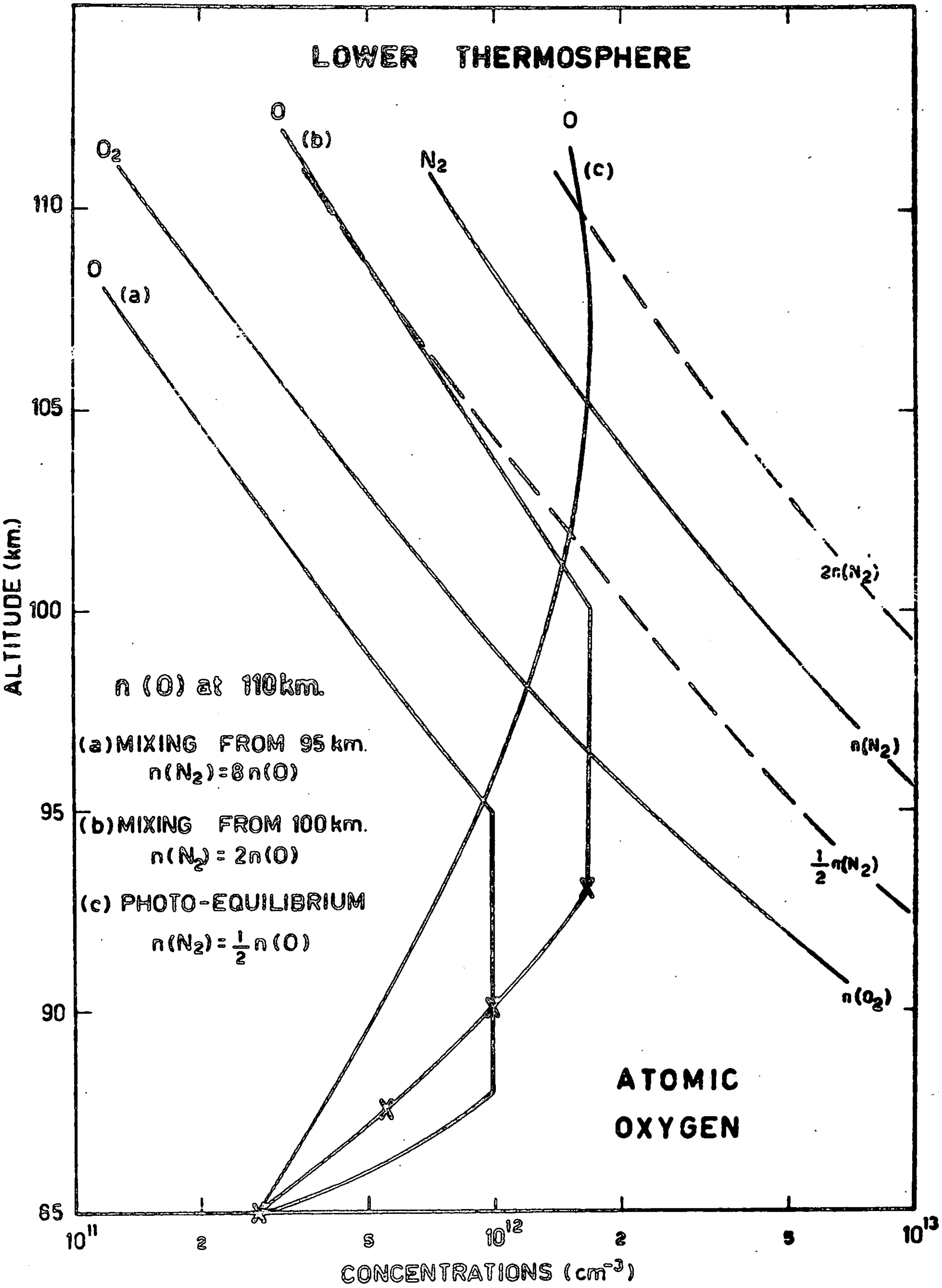
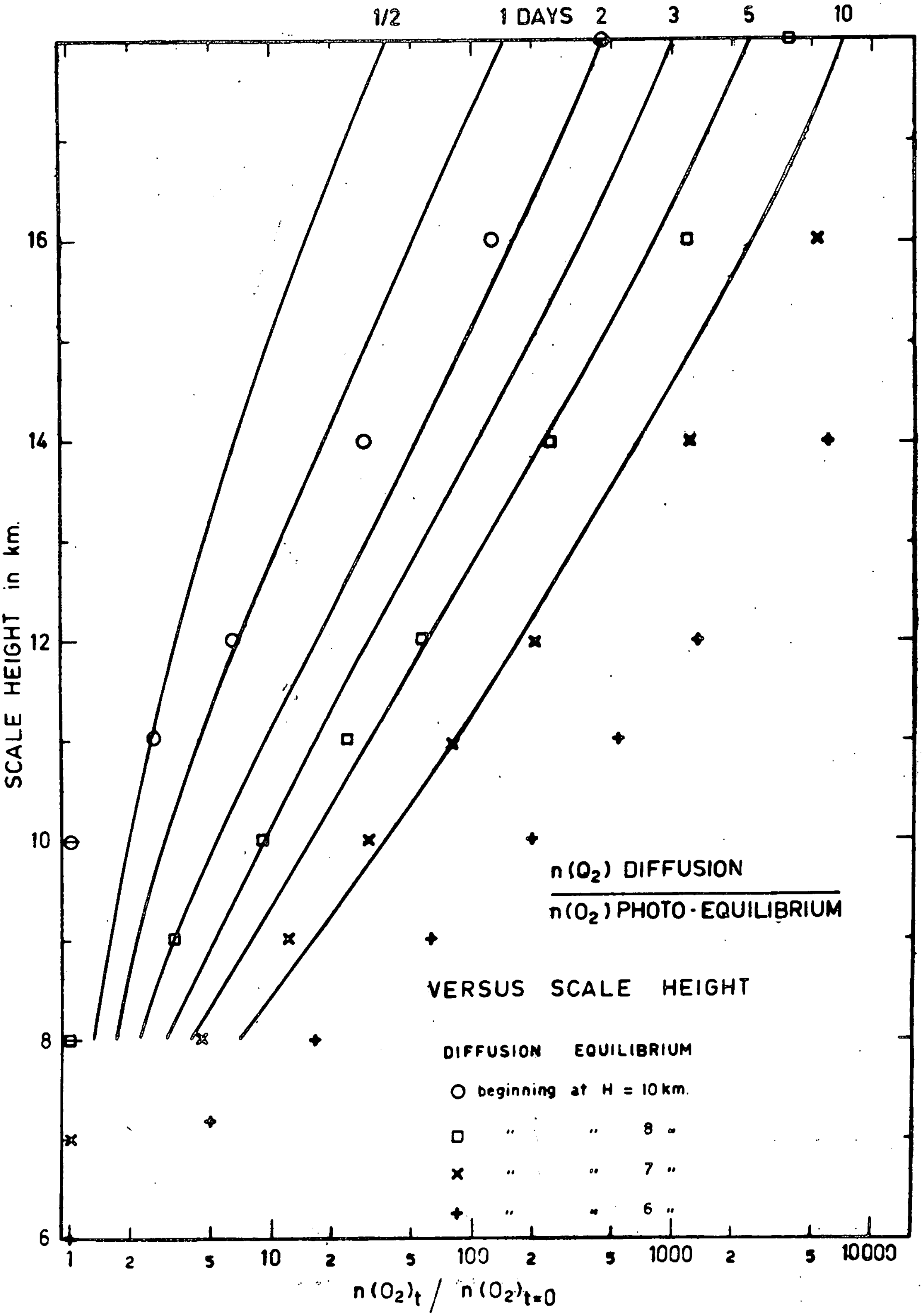

LOWER THERMOSPHERE
O₂
O (a)
O (b)
N₂
O (c)
110
105
100
95
90
85
ALTITUDE (km.)
n (O) at 110km.
(a) MIXING FROM 95 km. n(N₂)=8 n(O)
(b) MIXING FROM 100 km. n(N₂) = 2n(O)
(c) PHOTO-EQUILIBRIUM n(N₂) = 1/2 n(O)
2n(N₂)
n(N₂)
1/2 n(N₂)
n(O₂)
ATOMIC OXYGEN
10¹¹
2
5
10¹²
2
5
10¹³
CONCENTRATIONS (cm⁻³)

Figure 2.



1/2
1 DAYS
2
3
5
10
SCALE HEIGHT in km.
6
8
10
12
14
16
1
2
5
10
2
5
100
2
5
1000
2
5
10000
$n(O_2)_t \,/\, n(O_2)_{t=0}$
$\frac{n(O_2)\ \text{DIFFUSION}}{n(O_2)\ \text{PHOTO-EQUILIBRIUM}}$
VERSUS SCALE HEIGHT
DIFFUSION EQUILIBRIUM
○ beginning at H = 10 km.
□ " " 8 "
× " " 7 "
+ " " 6 "

Figure 3.

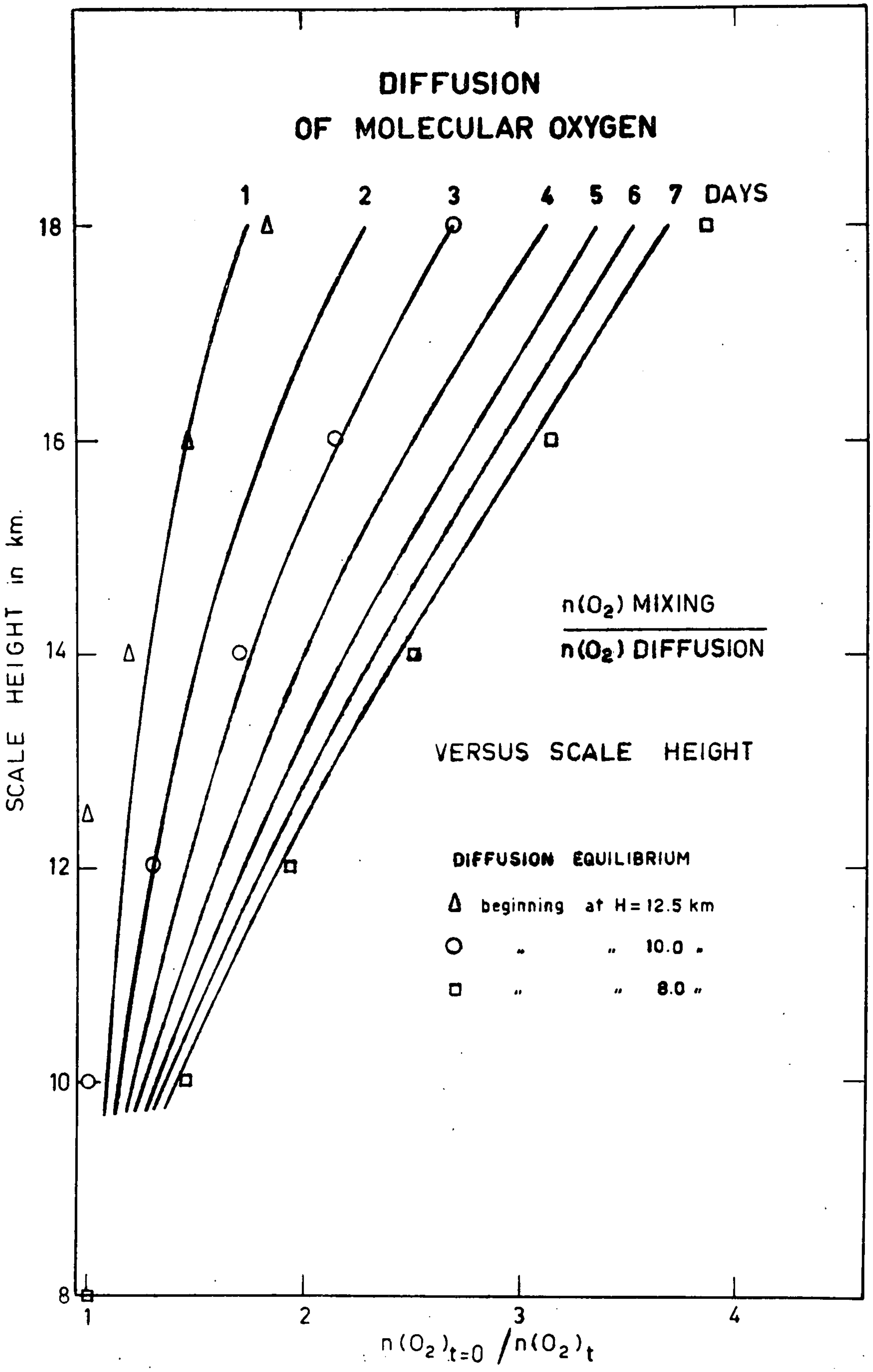

DIFFUSION
OF MOLECULAR OXYGEN
1
2
3
4
5
6
7
DAYS
18
16
14
12
10
8
SCALE HEIGHT in km.
1
2
3
4
$n(O_2)_{t=0} / n(O_2)_t$
$n(O_2)$ MIXING
$n(O_2)$ DIFFUSION
VERSUS SCALE HEIGHT
DIFFUSION EQUILIBRIUM
Δ beginning at H = 12.5 km
○ " " 10.0 "
□ " " 8.0 "

Figure 4.

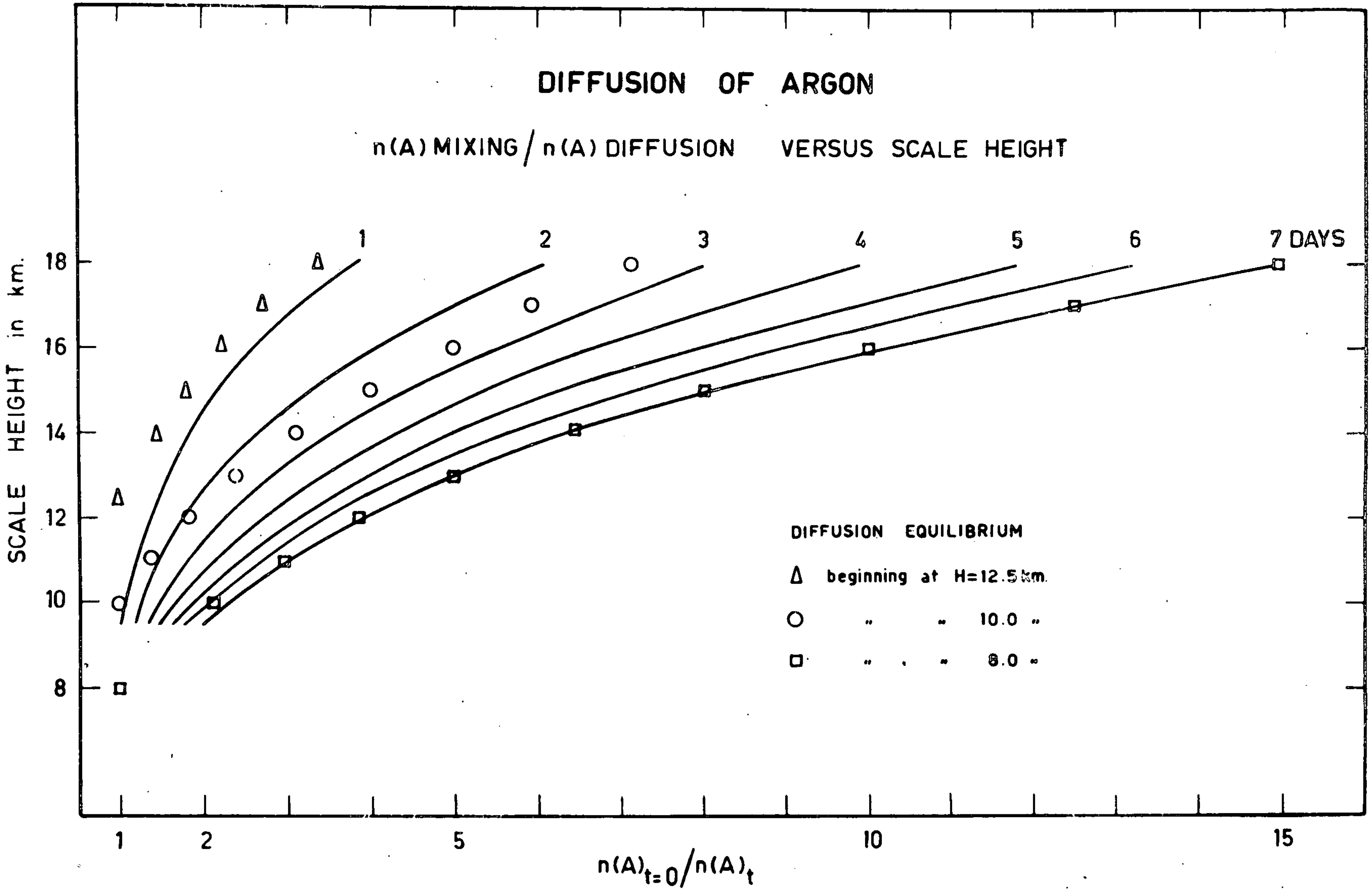

DIFFUSION OF ARGON
n(A) MIXING / n(A) DIFFUSION VERSUS SCALE HEIGHT
1
2
3
4
5
6
7 DAYS
SCALE HEIGHT in km.
18
16
14
12
10
8
DIFFUSION EQUILIBRIUM
Δ beginning at H=12.5 km.
○ " " 10.0 "
□ " " 8.0 "
1
2
5
10
15
$n(A)_{t=0}/n(A)_t$

Figure 5.

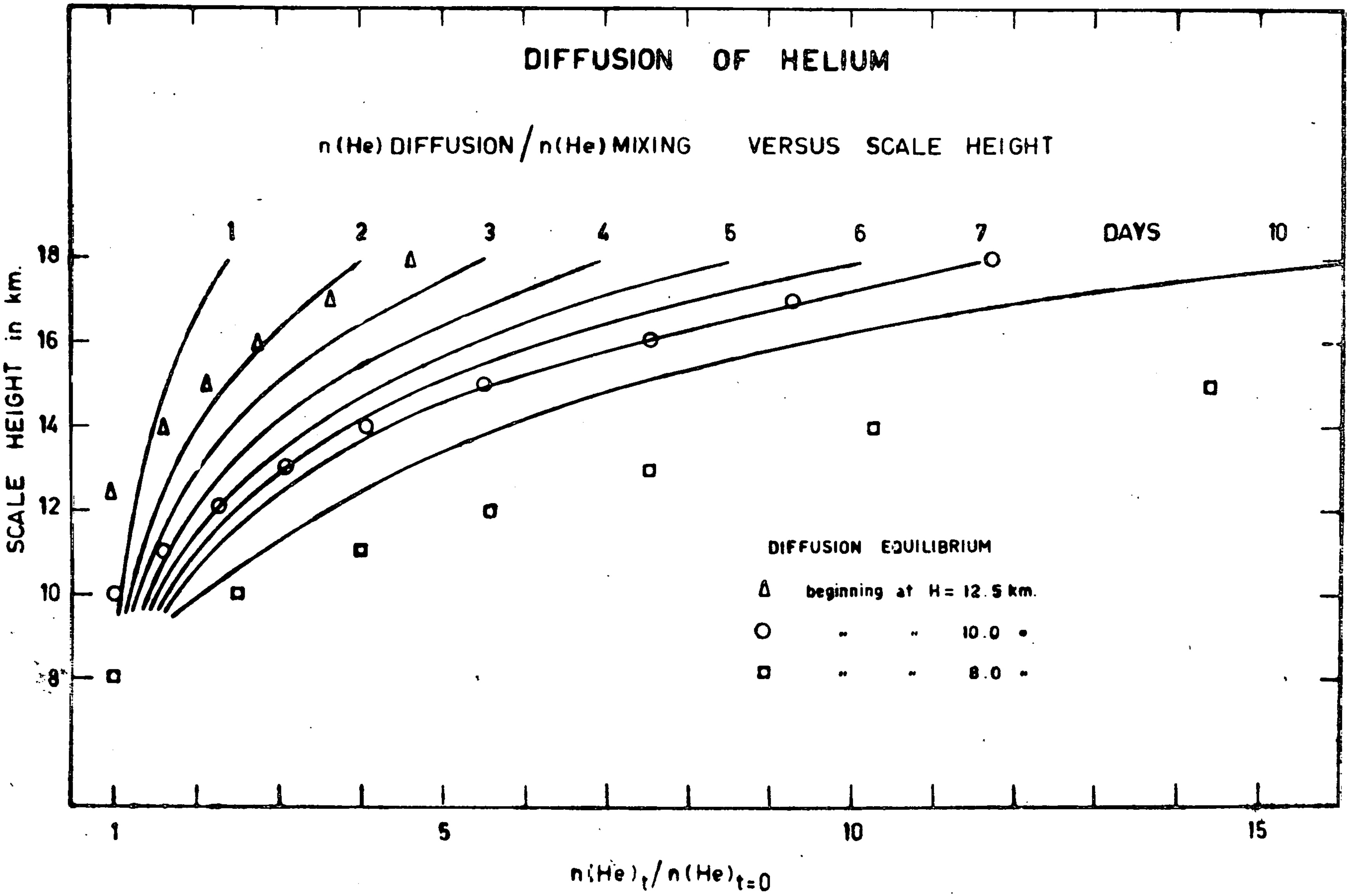

DIFFUSION OF HELIUM
n(He) DIFFUSION / n(He) MIXING VERSUS SCALE HEIGHT
1
2
3
4
5
6
7
DAYS
10
SCALE HEIGHT in km.
18
16
14
12
10
8
1
5
10
15
n(He)t / n(He)t=0
DIFFUSION EQUILIBRIUM
Δ beginning at H = 12.5 km.
○ " " 10.0 "
□ " " 8.0 "

Figure 6.

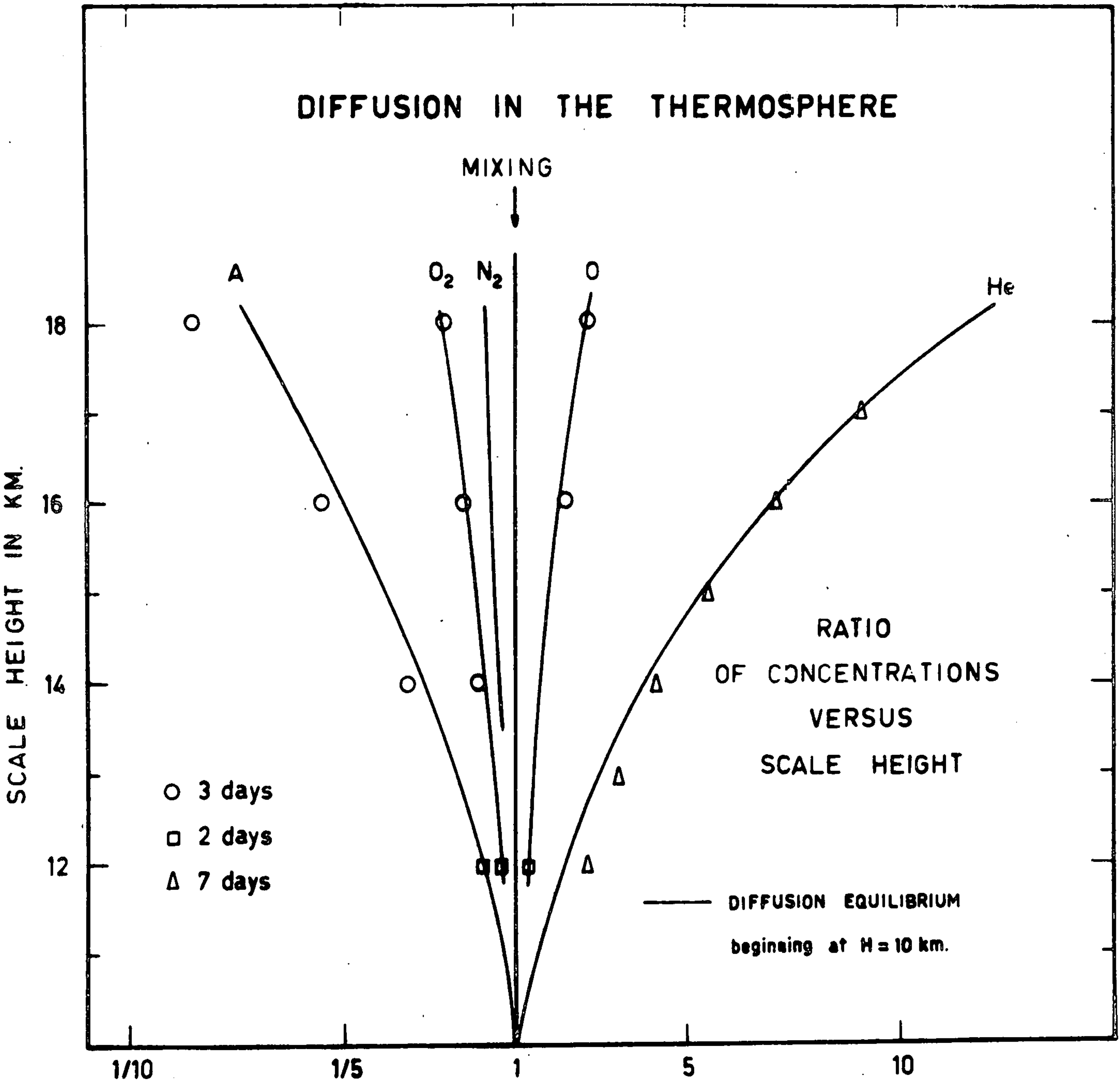

DIFFUSION IN THE THERMOSPHERE
MIXING
A
O2
N2
O
He
18
16
14
12
SCALE HEIGHT IN KM.
RATIO OF CONCENTRATIONS VERSUS SCALE HEIGHT
○ 3 days
□ 2 days
Δ 7 days
DIFFUSION EQUILIBRIUM beginning at H = 10 km.
1/10
1/5
1
5
10

Figure 7.

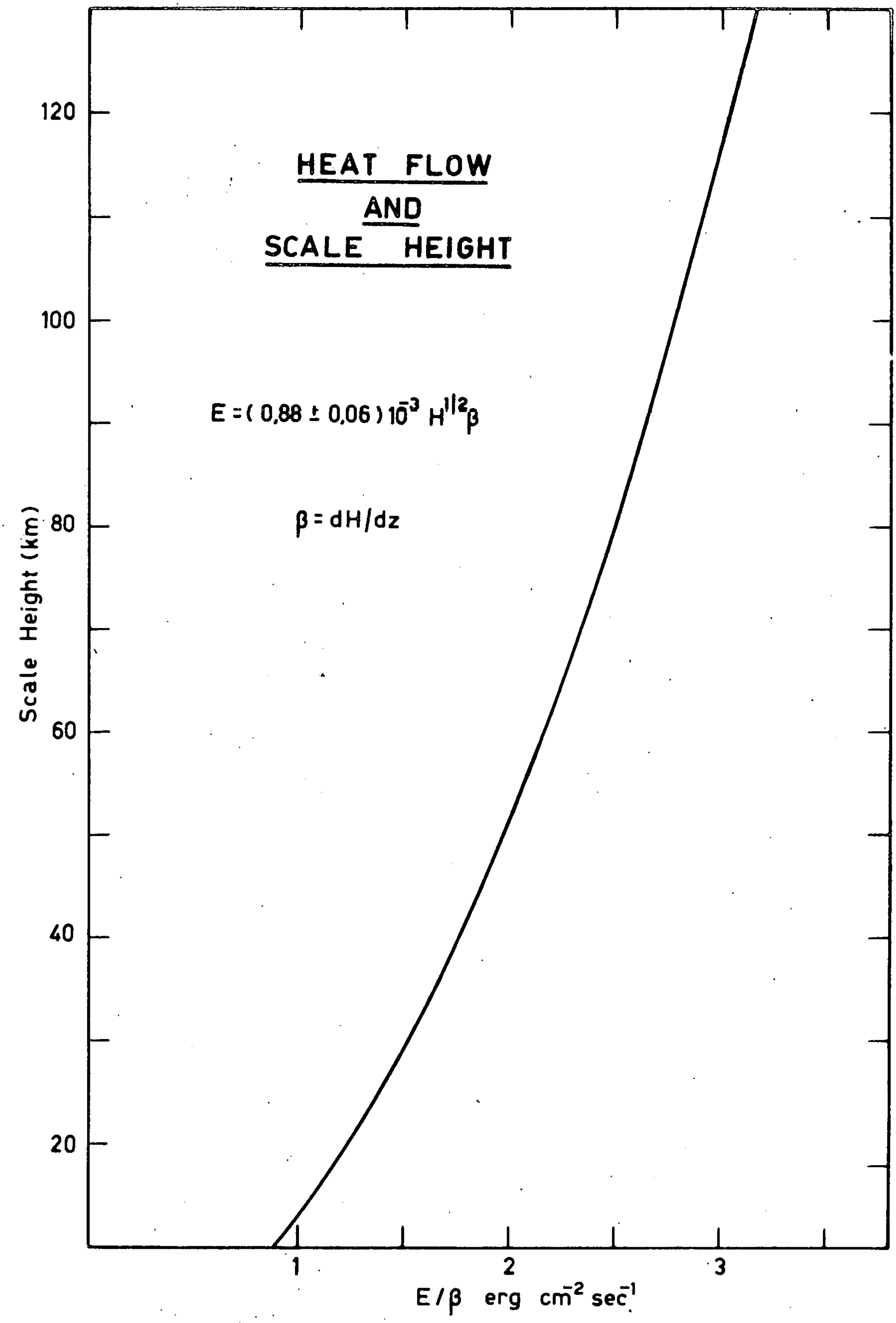

HEAT FLOW
AND
SCALE HEIGHT
E = (0,88 ± 0,06) 10⁻³ H^1/2 β
β = dH/dz
Scale Height (km)
120
100
80
60
40
20
1
2
3
E/β erg cm⁻² sec⁻¹

Figure 8.

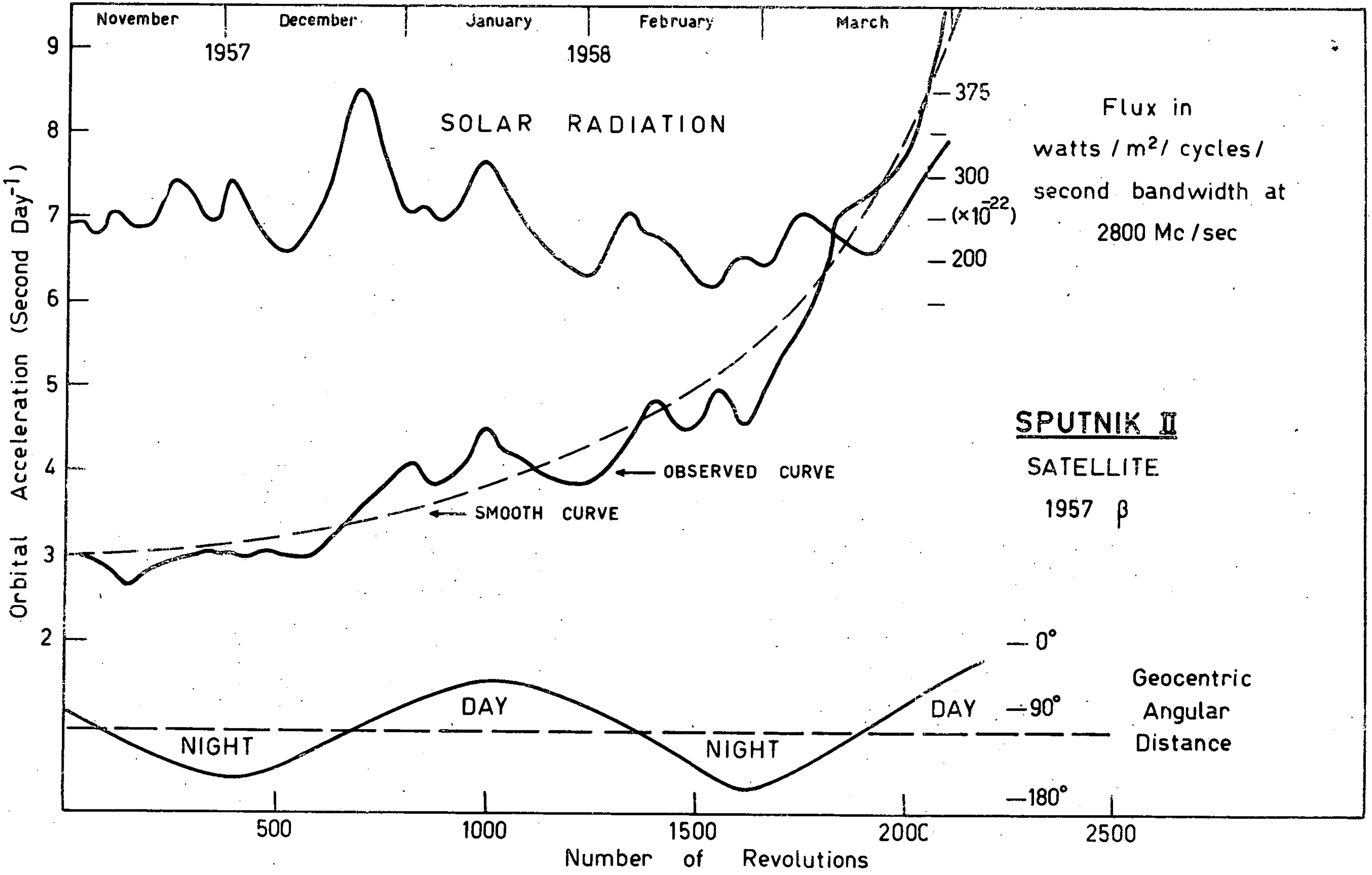

November
December
January
February
March
1957
1958
9
8
7
6
5
4
3
2
Orbital Acceleration (Second Day^{-1})
SOLAR RADIATION
375
300
$(\times 10^{-22})$
200
Flux in watts / m²/ cycles / second bandwidth at 2800 Mc / sec
SPUTNIK II
SATELLITE
1957 β
OBSERVED CURVE
SMOOTH CURVE
0°
90°
180°
DAY
NIGHT
DAY
NIGHT
Geocentric Angular Distance
500
1000
1500
2000
2500
Number of Revolutions

Figure 9.

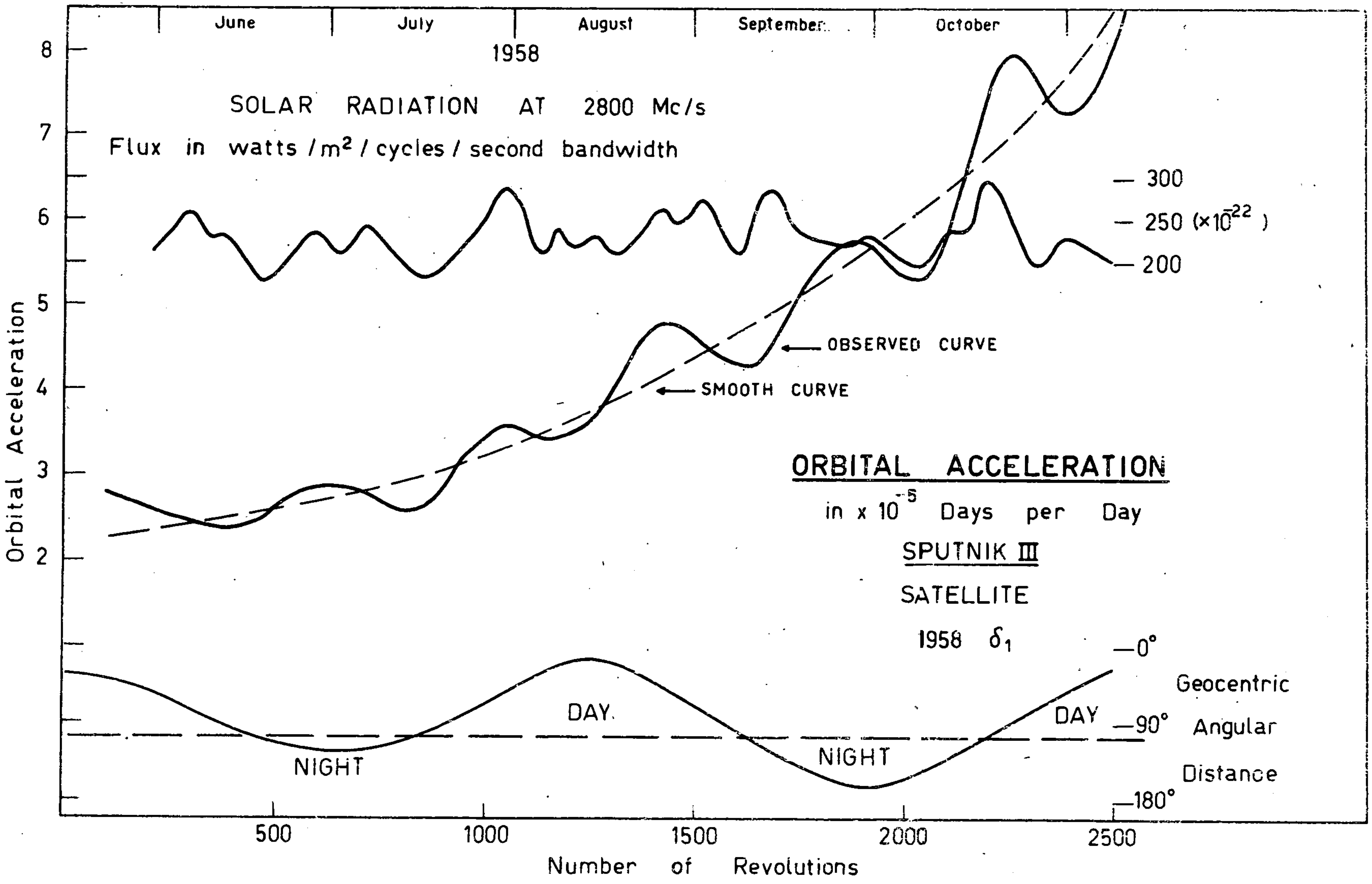

June
July
August
September
October
1958
SOLAR RADIATION AT 2800 Mc/s
Flux in watts / m² / cycles / second bandwidth
300
250 (×10⁻²²)
200
OBSERVED CURVE
SMOOTH CURVE
ORBITAL ACCELERATION
in x 10⁻⁵ Days per Day
SPUTNIK III
SATELLITE
1958 δ1
0°
90°
180°
Geocentric Angular Distance
DAY
NIGHT
DAY
NIGHT
Orbital Acceleration
8
7
6
5
4
3
2
500
1000
1500
2000
2500
Number of Revolutions

Figure 10 a.

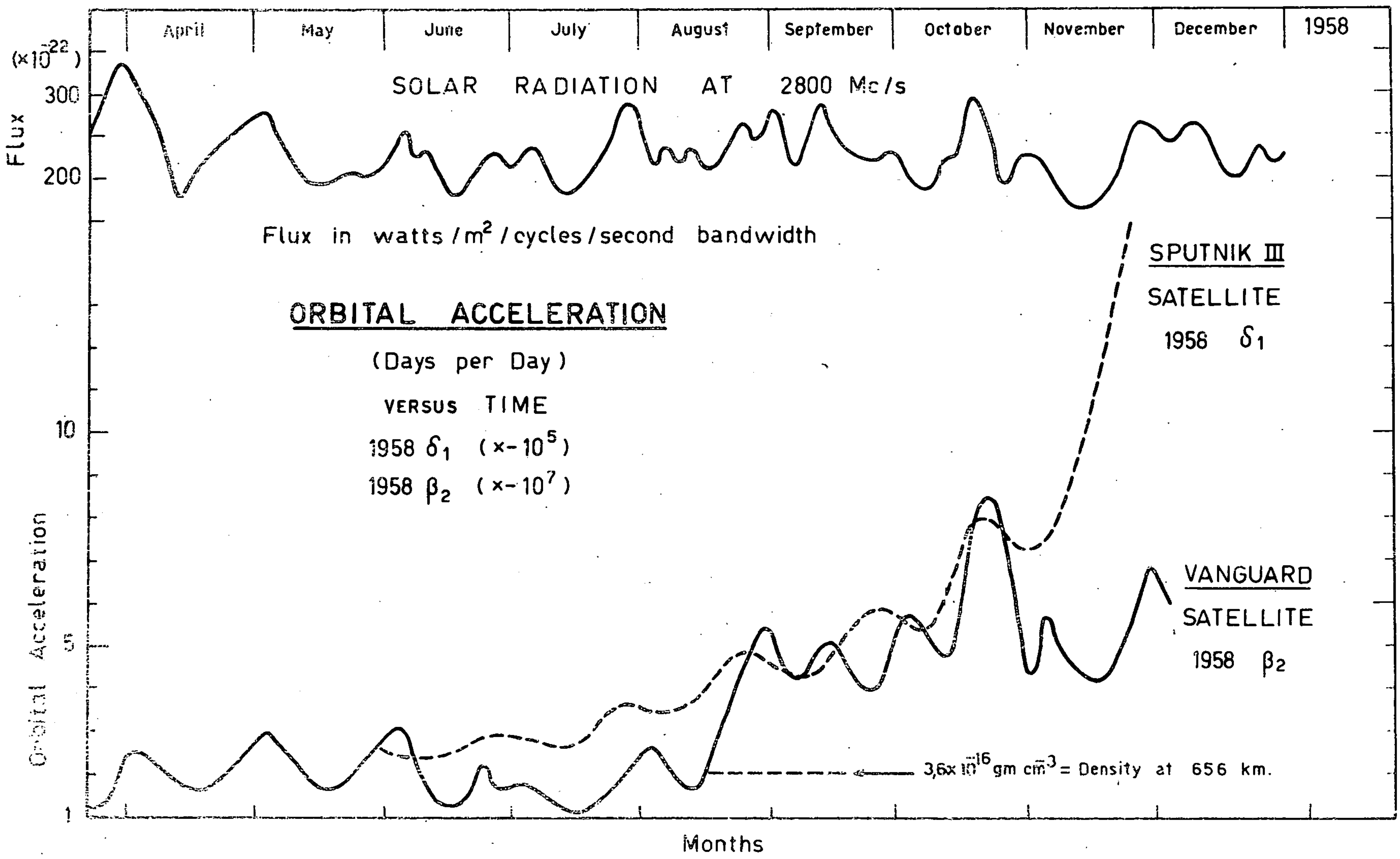
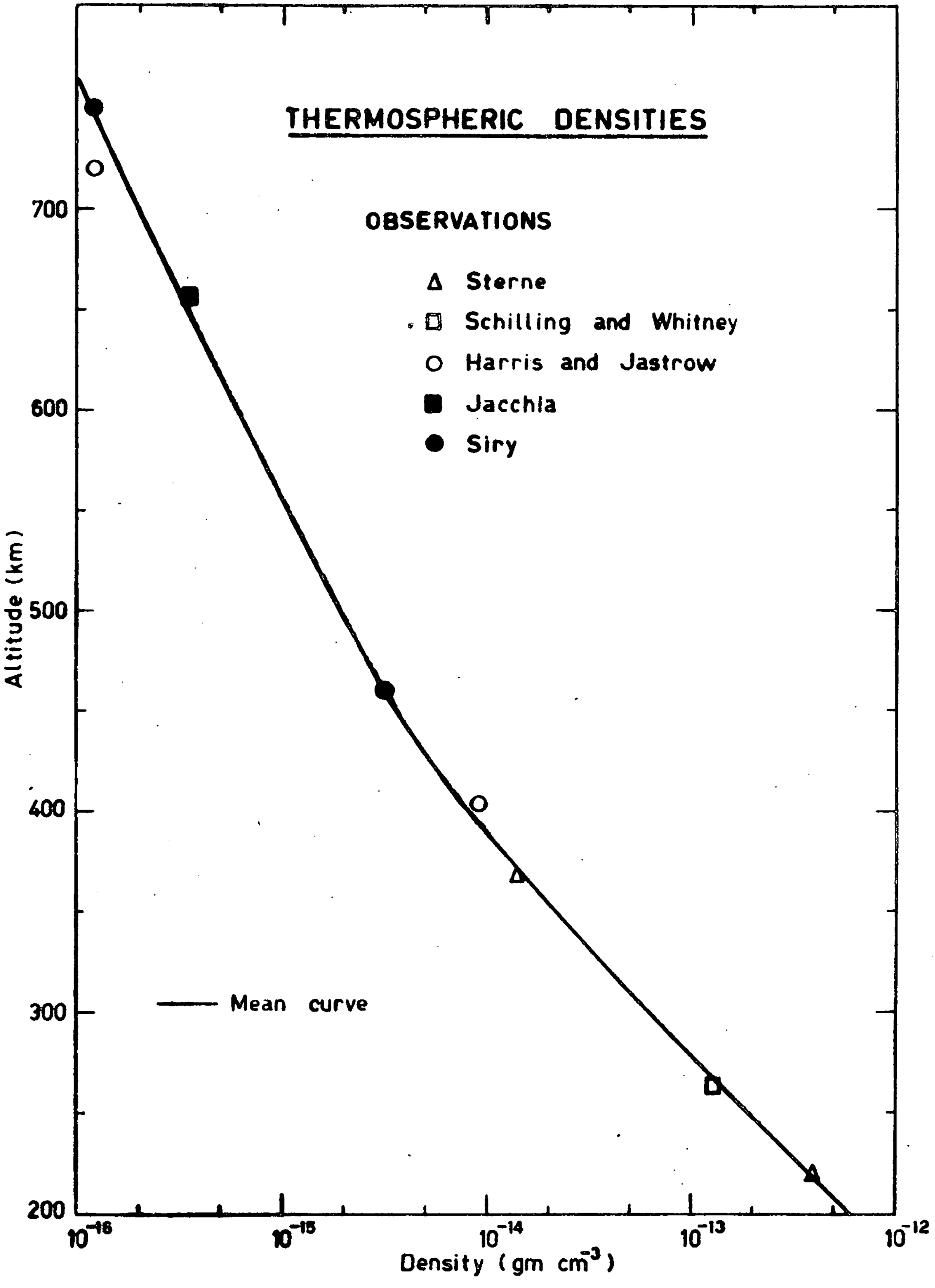

April
May
June
July
August
September
October
November
December
1958
(×10^{-22})
300
200
Flux
SOLAR RADIATION AT 2800 Mc/s
Flux in watts/m^2/cycles/second bandwidth
ORBITAL ACCELERATION
(Days per Day)
VERSUS TIME
1958 δ_1 (×-10^5)
1958 β_2 (×-10^7)
SPUTNIK III
SATELLITE
1958 δ_1
VANGUARD
SATELLITE
1958 β_2
3,6x10^{-16} gm cm^{-3} = Density at 656 km.
10
5
1
Orbital Acceleration
Months

Figure 10 b.



THERMOSPHERIC DENSITIES
OBSERVATIONS
△ Sterne
□ Schilling and Whitney
○ Harris and Jastrow
■ Jacchia
● Siry
Mean curve
Altitude (km)
700
600
500
400
300
200
10^{-16}
10^{-15}
10^{-14}
10^{-13}
10^{-12}
Density (gm cm^{-3})

Figure 11.

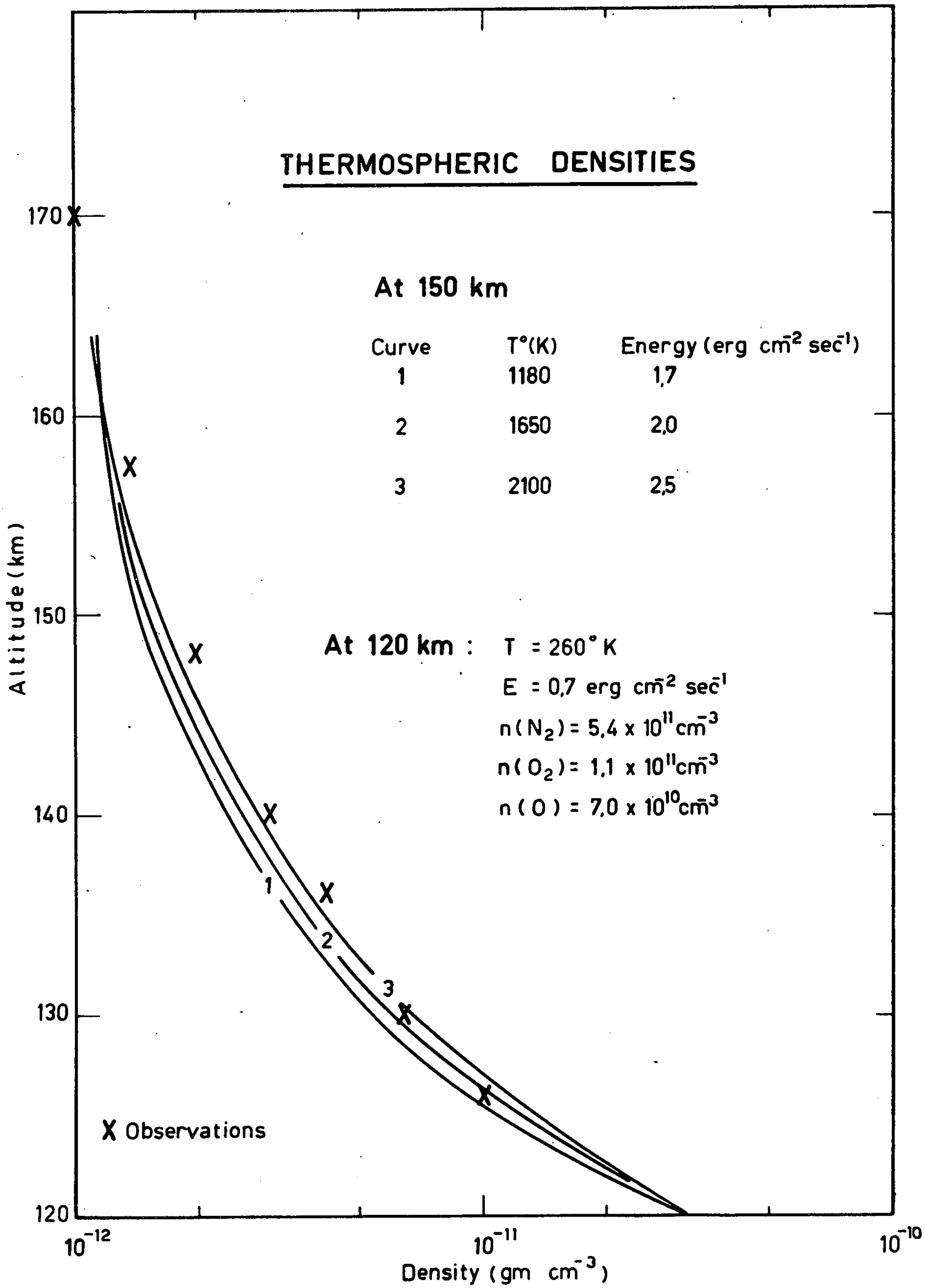
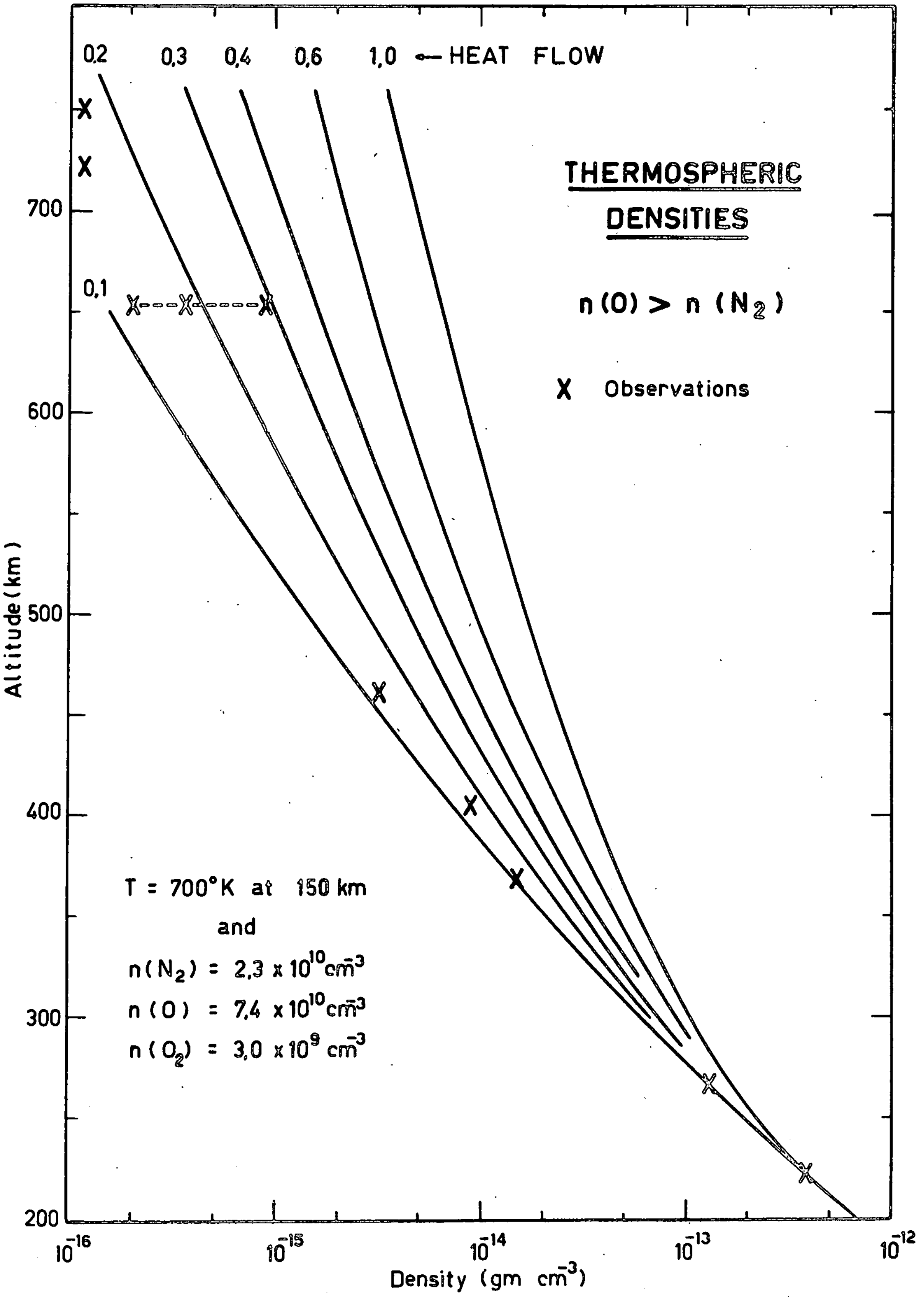

THERMOSPHERIC DENSITIES
At 150 km
Curve 1: T°(K) 1180, Energy (erg cm^{-2} sec^{-1}) 1,7
Curve 2: T°(K) 1650, Energy (erg cm^{-2} sec^{-1}) 2,0
Curve 3: T°(K) 2100, Energy (erg cm^{-2} sec^{-1}) 2,5
At 120 km : T = 260° K
E = 0,7 erg cm^{-2} sec^{-1}
$n(N_2) = 5{,}4 \times 10^{11}$ cm^{-3}
$n(O_2) = 1{,}1 \times 10^{11}$ cm^{-3}
$n(O) = 7{,}0 \times 10^{10}$ cm^{-3}
X Observations
1
2
3
Altitude (km)
120
130
140
150
160
170
Density (gm cm^{-3})
10^{-12}
10^{-11}
10^{-10}

Figure 12.



0,2
0,3
0,4
0,6
1,0
HEAT FLOW
0,1
THERMOSPHERIC DENSITIES
n(O) > n (N₂)
X Observations
T = 700°K at 150 km
and
n(N₂) = 2,3 x 10¹⁰cm⁻³
n (O) = 7,4 x 10¹⁰cm⁻³
n (O₂) = 3,0 x 10⁹ cm⁻³
Altitude (km)
700
600
500
400
300
200
10⁻¹⁶
10⁻¹⁵
10⁻¹⁴
10⁻¹³
10⁻¹²
Density (gm cm⁻³)

Figure 13.

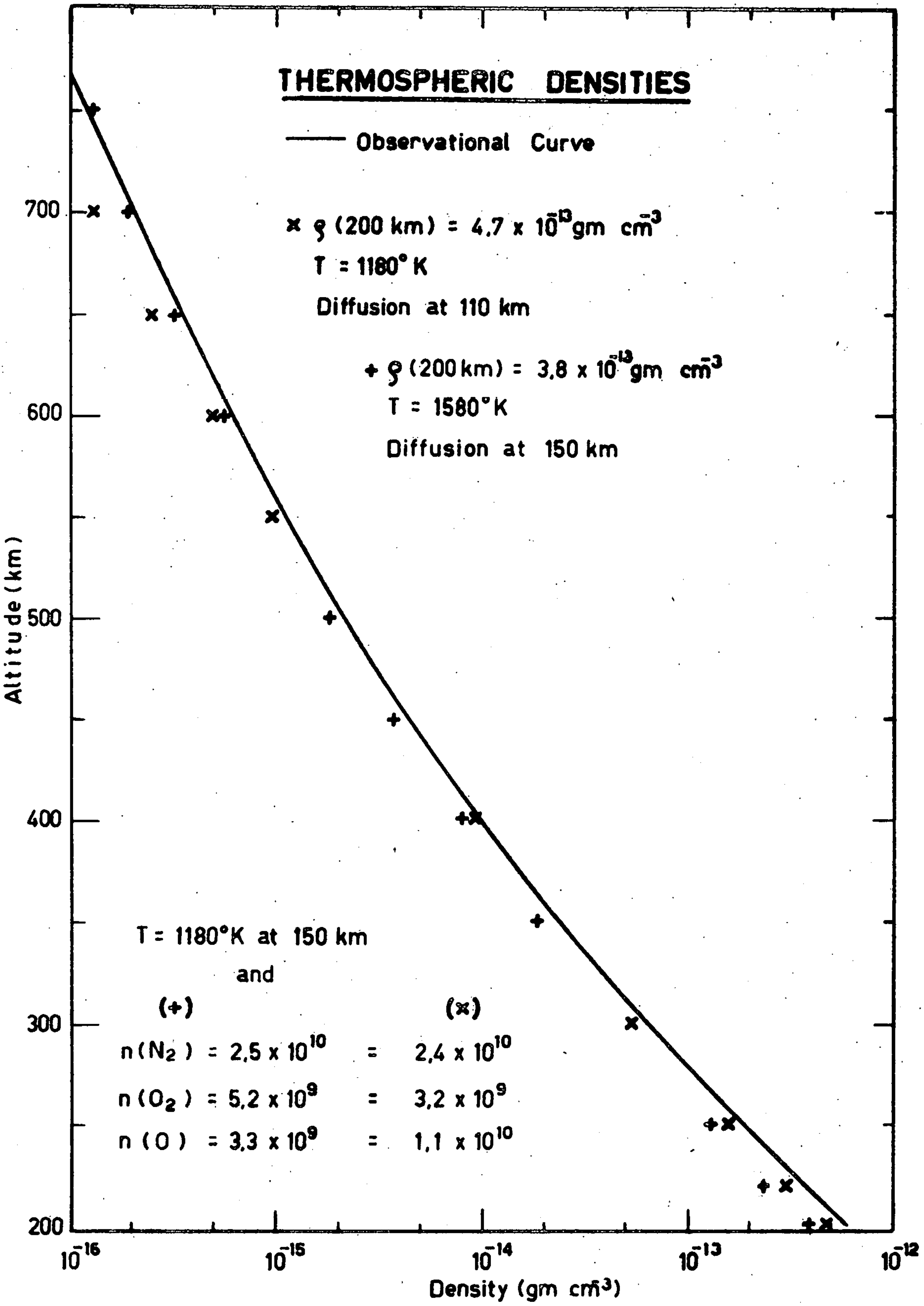
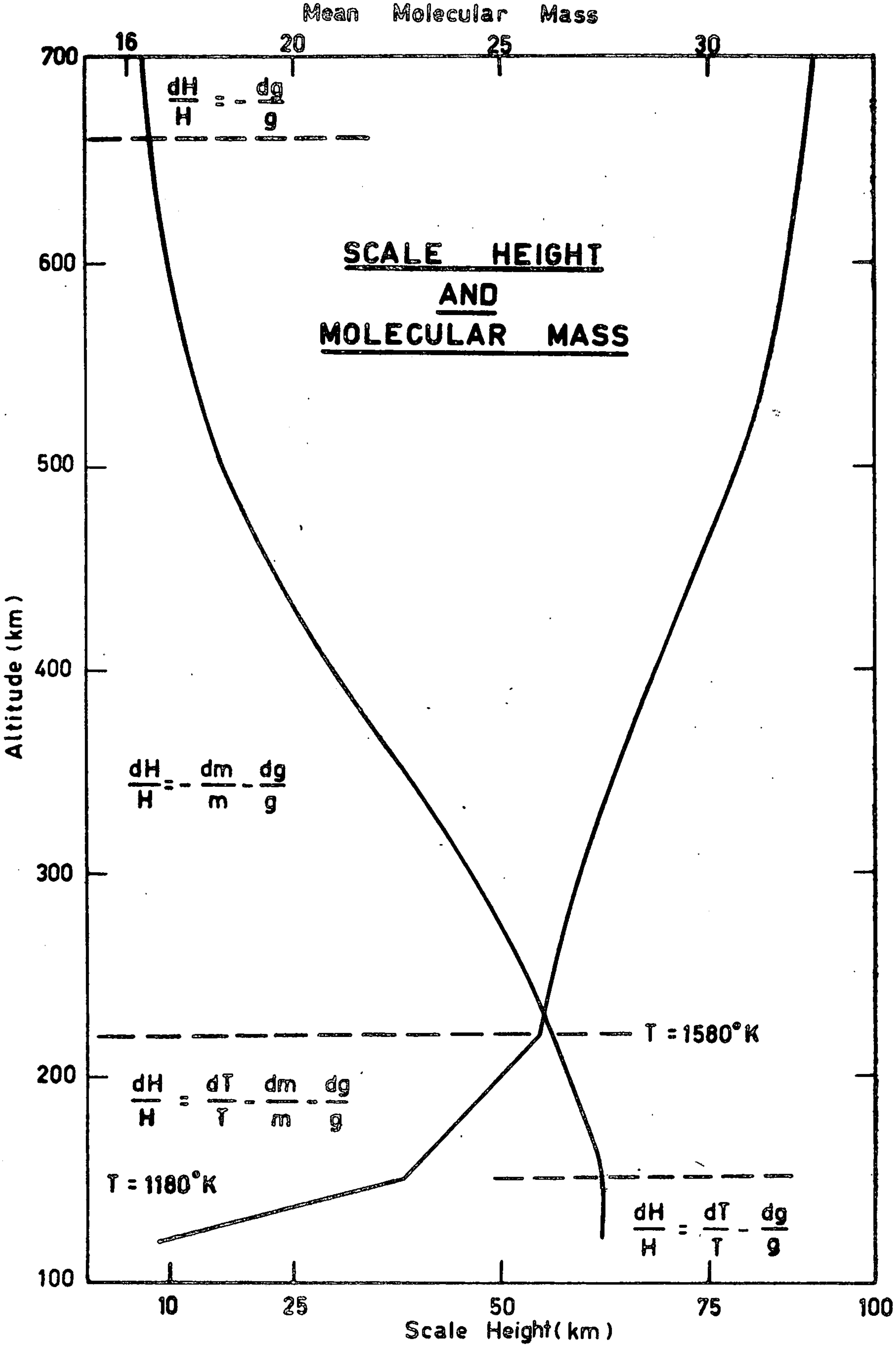

THERMOSPHERIC DENSITIES
Observational Curve
× ϱ (200 km) = 4,7 x 10⁻¹³ gm cm⁻³
T = 1180° K
Diffusion at 110 km
+ ϱ (200 km) = 3,8 x 10⁻¹³ gm cm⁻³
T = 1580° K
Diffusion at 150 km
T = 1180°K at 150 km
and
(+) (×)
n(N2) = 2,5 x 10^10 = 2,4 x 10^10
n(O2) = 5,2 x 10^9 = 3,2 x 10^9
n(O) = 3,3 x 10^9 = 1,1 x 10^10
Altitude (km)
700
600
500
400
300
200
10^-16
10^-15
10^-14
10^-13
10^-12
Density (gm cm⁻³)

Figure 14.



Mean Molecular Mass
16
20
25
30
700
600
500
400
300
200
100
Altitude (km)
10
25
50
75
100
Scale Height (km)
SCALE HEIGHT AND MOLECULAR MASS
$\frac{dH}{H} = -\frac{dg}{g}$
$\frac{dH}{H} = -\frac{dm}{m} - \frac{dg}{g}$
T = 1580°K
$\frac{dH}{H} = \frac{dT}{T} - \frac{dm}{m} - \frac{dg}{g}$
T = 1180°K
$\frac{dH}{H} = \frac{dT}{T} - \frac{dg}{g}$

Figure 15.

REFERENCES

1. M. Nicolet, Annales Géophys. 13, 1(1957).
2. E. Gluckauf, in 'Compendium of Meteorology', p.1, Amer. Meteo. Soc. Boston, Mass. 1951.
3. D.R. Bates, in 'The Earth as a Planet' (G.P. Kuiper, ed), Chapter 12, University of Chicago Press, 1954.
4. F.A. Paneth, J. Chem. Soc., Pt III, 3651 (1952)
5. E. Gluckauf and F.A. Paneth, Proc. Roy. Soc. A 185,69,1946.
6. J.W. Townsend, Jr., E.B. Meadows, and E.C. Pressly, in 'Rocket Exploration of the Upper Atmosphere' (R.L.F. Boyd and M.J. Seaton, ed.), p. 169, Pergamon Press, London,1954.
7. E.B. Meadows, and J.W. Townsend, Jr., Annales Géophys. 14, 80 (1958)
8. E.T. Byram, T.A. Chubb and H. Friedman, in 'The Proceedings of the Conference on Chemical Aeronomy, (M. Zelikoff, ed.) p. 211, Pergamon Press, London, 1957.
9. M. Nicolet and P. Mange, J. Geophys. Res., 59, 16 (1954)
10. M. Nicolet, J. Atm. Terr. Phys., 5, 132 (1954)
11. D.R. Bates and M. Nicolet, J. Geophys. Res., 55, 301 (1950)
12. M. Nicolet, Annales Géophys. 14, - (1958)
13. S. Chapman and T.G. Cowling, 'The Mathematical Theory of Non-Uniform Gases', Cambridge University Press 1939 and 1952.
14. M. Nicolet, in 'The Earth as a Planet' (G.P. Kuiper, ed.), Chapter 13, University of Chicago Press, 1954.

15. P.S. Epstein, Beitr. Geophys., 35, 153 (1932)

16. P. Mange, Annales Géophys. 11, 153 (1955)

17. P. Mange, J. Geophys. Res., 62, 279 (1957)

18. A. Dalgarno, J. Atm. Terr. Phys., 12, 219 (1958)

19. V.C. Ferraro and I. Ozdogan, J. Atm. Terr. Phys., 12, 140 (1958)

20. L. Spitzer, Jr., in 'The Atmospheres of the Earth and Planets', (G.P. Kuiper, ed.) p. 211, University of Chicago Press, 1952.

21. D.R. Bates, Proc. Phys. Soc., B 64, 805 (1951).

22. D.R. Bates, Proc. Phys. Soc., A. 236, 206 (1956).

23. F.S. Johnson, J. Geophys. Res., 61, 71 (1956).

24. S. Chapman, Smithsonian Cont. Astrophys., 2, 1 (1957).

25. A.N. Lowan, J. Geophys. Res., 60, 421 (1955).

26. L. Oster, Zeits. Astrophys., 40, 28 (1956).

27. J. Kestin, in 'American Institute of Physics Handbook' (D.E. Gray, ed.), section 2, p 201, McGraw-Hill, New York (1957).

28. R.J. Havens, R.T. Koll, and H.E. LaGow, J. Geophys. Res., 57, 59 (1952).

29. The Rocket Panel, Phys. Rev., 88, 1027 (1952).

30. F.L. Whipple, in 'The Earth as a Planet' (G.P. Kuiper, ed.), Chapter 10, University of Chicago Press, 1954.

31. W.G. Stroud, W. Nordberg and J.R. Walsh, J. Geophys. Res., 61, 45 (1956).

32. H.E. LaGow, R. Horowitz and J. Ainsworth, Annales Géophys., 14, 131 (1958).

33. L.M. Jones, F.F. Fischbach and J.W. Peterson, IGY Rocket Report Series, Washington n°1, 47 (1958).

34. R. Horowitz, and H.E. LaGow, J. Geophys. Res., 62, 57 (1957).

35. E.T. Byram, T.A. Chubb, and H. Friedman, J. Geophys. Res., 61, 251 (1956).

36. V.V. Mikhnevitch, Progress in Physical Sciences, Acad. Sc. USSR, 63, 197 (1957).

37. H.E. LaGow, R. Horowitz and J. Ainsworth, IGY Rocket Report Series, Washington, n°1, 38 (1958).

38. J.E. Kupperian Jr., E.T. Byram, H. Friedman and A. Unzicker, IGY Rocket Report Series, Washington, n°1, 203 (1958).

39. J.W. Townsend, Jr., and E.B. Meadows, Annales Géophys., 14, 117 (1958).

40. T.E. Sterne, Science, 128, 420 (1958).

41. T.E. Sterne, Phys. Fluids, 1, 165 (1958).

42. T.E. Sterne and G.F. Schilling, Smith. Cont. Astrophys., 2, 207 (1958).

43. G.F. Schilling and T.E. Sterne, J. Geophys. Res., 40, n°1 (1959).

44. I. Harris and R. Jastrow, Science, 127, 471 (1958).

45. I. Harris and R. Jastrow, Science, 128, 421, (1958).

46. L.G. Jacchia, Smiths. Astrophys. Obs., Special Report n°13, (1958).

47. L.G. Jacchia, Smiths. Astrophys. Obs., Special Report nº 12, 30 (1958).

48. L.G. Jacchia and R.E. Briggs, Smiths. Astrophys. Obs., Special Report nº 18, 9 (1958).

49. G.F. Schilling and C.A. Whitney, Smiths. Astrophys. Obs., Special Report nº 18, 13 (1958).

50. H.K. Paetzold and H. Zschörner, Die Naturwissenschaften, 45, 485 (1958).

51. G.V. Groves, Nature, 181, 1055 (1958).

52. D.G. King-Hele and D.C. Leslie, Nature, 182, 860 (1958).

53. C.H. Bosanquet, Nature, 182, 1533 (1958).

54. G.V. Groves, Nature, 182, 1533 (1958).

55. J.W. Siry, to be published in Annals of IGY.

56. M. Nicolet, Science, 127, 1317 (1958).

57. R.M. Merson, D.G. King-Hele and R.N.A. Plimmer, Nature, 183, 239 (1959).

58. J.W. Townsend, Jr., Science, 129, 80 (1959).