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ATMOSPHERE

by

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containing careful surveys, maps, and sketches; J. THOMSON, *Travels in the Atlas and Southern Morocco* (1889), includes important botanical and geological data; W.B. HARRIS, *Tafilet: The Narrative of a Journey of Exploration in the Atlas Mountains and the Oases of the North-West Sahara* (1895); G.H. BOUSQUET, *Les Berbères*, 3rd ed. (1967), describes the people, their institutions, customs, and religion; J. BERQUE, *Structures sociales du Haut-Atlas* (1955). For the Atlas Tellien, see X. YACONO, *La Colonisation des plaines du Chélif de Lavigerie confluent de la Mina*, 2 vol. (1955–56); and *Les Bureaux arabes et l'évolution des genres de vie indigènes dans l'ouest du Tell algérois* (1953), which covers Dahra, Chélif, Ouarsenis, and Sersou. For the Atlas Saharien, see R. FURON, *Le Sahara* (1957), for geological data; and R. CAPOT-REY, *Le Sahara français* (1953).

(H.Is.)

Atmosphere

The atmosphere that surrounds the Earth and is commonly called the air consists of layers of gases and mixtures of gases, as well as water vapour and solid and liquid particles. The mean pressure exerted by a vertical column of the atmosphere at sea level equals that of a column of mercury 760 millimetres (29.92 inches) in height. Such a column of air exerts a force, called atmospheric pressure, expressed as 1.033 kilograms per square centimetre (14.7 pounds per square inch).

Higher in the atmosphere the pressure decreases, and at an altitude of six kilometres (four miles) it is only one-half that at sea level. If this decrease in the pressure within the field of gravity were uniform throughout the depth of the atmosphere, the weight of a column of air at about 60 kilometres (40 miles) would be reduced to a thousandth (10^{-3}) that at sea level, to a thousandth of a millionth (10^{-9}) at about 180 kilometres (110 miles), and to 10^{-21} at about 420 kilometres (260 miles). The last value (10^{-21}) represents a density less than that of interstellar space. It is known, however, that the atmosphere of the Earth extends well beyond 400 kilometres (250 miles). Visual observations—of the auroras, for example—reveal the presence of luminous rays up to altitudes as great as 1,000 kilometres (600 miles). Furthermore, shortwave radio transmission has proved that the medium is sufficiently dense at an altitude of several hundred kilometres to produce enough electric charges (electrons) to reflect radio waves. Rocket probes and especially the drag encountered by artificial satellites at altitudes of several thousand kilometres have demonstrated that the terrestrial atmosphere extends to a very great distance. This extension to high altitudes occurs because of the occurrence of constituents of low mass and because of pressure and temperature conditions.

This article treats the physical and chemical properties of the several regions of the atmosphere. For further information on the interaction of the Earth's magnetic field and the atmosphere, see EARTH, MAGNETIC FIELD OF; IONOSPHERE; VAN ALLEN RADIATION BELTS; and AURORAS. See ATMOSPHERE, DEVELOPMENT OF for treatment of the origin and evolution of the present atmosphere; and WINDS AND STORMS; OCEANS AND SEAS; and CLIMATE for the role of the atmosphere at levels near the Earth and its interaction with the hydrosphere.

REGIONS OF THE ATMOSPHERE

Vertical temperature distribution

The principal regions of the atmosphere, each of which is characterized by the pattern of vertical distribution of temperatures, are the troposphere, the stratosphere, the mesosphere, the thermosphere, and the exosphere (Figure 1). In meteorological research it has long been customary to deal with two regions: the troposphere and the stratosphere.

The lower atmosphere. *The troposphere.* The region of the atmosphere in contact with the Earth's surface, the troposphere, is the realm of the clouds, rain, snow, etc., and is characterized in general by a decrease of temperature with increasing altitude. The upper limit of the troposphere, known as the tropopause, is at an altitude of about 17 kilometres (11 miles) at the Equator and only six to eight kilometres (four to five miles) at the poles. In the middle latitudes, the altitude of this limit varies with

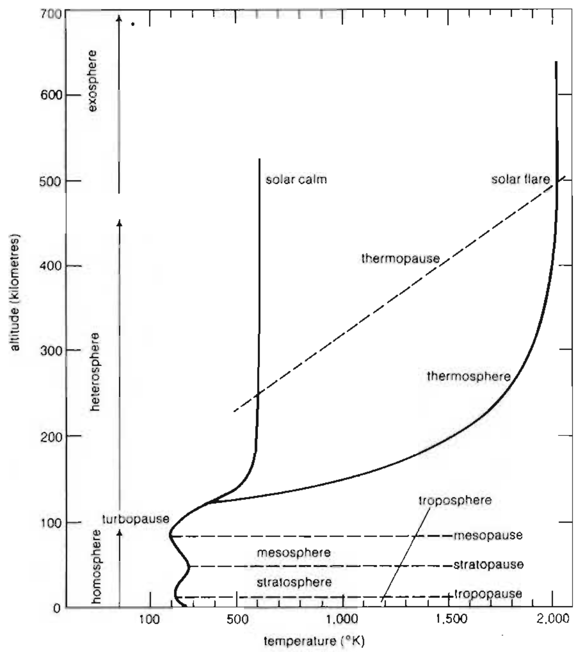


Figure 1: Regions of the atmosphere and their transitional zones.

the atmospheric conditions. In high-pressure areas it is about 13 kilometres (eight miles), and it may be below seven kilometres (four miles) under low-pressure conditions.

The stratosphere and mesosphere. The second region of the atmosphere that has been closely studied, the stratosphere, is still within the reach of meteorological probes. In its lower section there is a slow, constant increase in temperature with altitude; the temperature rise becomes more rapid with increasing altitude, attaining a maximum of about 270° K (the Kelvin scale is an absolute Celsius scale; *i.e.*, 0° K equals −273° C [−460° F], which is absolute zero) at approximately 50 kilometres (30 miles), the upper limit of the stratosphere, which is known as the stratopause. The temperature of the stratopause varies less than that of the tropopause, which may decrease from 220° K at the pole to about 190° K at the Equator. There is some variation of the temperature of the stratopause with latitude and with the seasons. The troposphere and the stratosphere are clearly separated; the exchange between tropospheric air and stratospheric air requires several months, if not several years, indicating a difference in the circulation of the air within the two zones.

Above 50 kilometres is the mesosphere or middle region, characterized by a rapid decrease of temperature to a minimum at about 85 kilometres (55 miles); this can be below 160° K in the summer at high latitudes. But, like the troposphere, the mesosphere is subject to strong seasonal variations of temperature at high latitudes. The prevailing level of the minimum temperature indicates another division in the atmosphere, known as the mesopause. The middle atmosphere

The upper atmosphere. Beyond the mesopause is an atmospheric region different in character from that of the lower regions. The first stratum of this higher region is the thermosphere, characterized by a continuous increase of temperature up to 500° K in the course of a night during minimum solar activity and to above 1,750° K in the course of a day during maximum solar activity. The altitude at which this increase of temperature ceases is the thermopause, which is at the base of an isothermal (constant-temperature) region that would extend into interplanetary space if the normal properties of a gas in hydrostatic (equal-sided) equilibrium continued to apply. In reality, the frequency of collisions between gas atoms becomes so low above a certain level, called the critical level, or baropause, or exobase, that the gas atoms can be considered to have their free-space trajectories. In this highest region, the exosphere (outersphere), the study of the physical properties becomes the study of the movement of particles subject to gravity and capable of escaping from the atmosphere. In the exosphere, temperature no longer has the customary meaning.

Atmospheric mixtures. Of the regions described above in terms of their vertical temperature distribution, the first three—troposphere, stratosphere, and mesosphere—have the same general composition; that is, the tropospheric mixture of molecular nitrogen and oxygen is maintained in the stratosphere and the mesosphere. In combination the three regions comprise the homosphere. At the higher altitude of 100 kilometres (60 miles), in the thermosphere, molecular oxygen is strongly dissociated, and atomic oxygen becomes an important component of the atmosphere. The latter region is known as the heterosphere and is characterized also by the presence of light atoms, such as helium and hydrogen, at its higher altitudes. A relative increase of these light elements in comparison with heavier elements such as nitrogen and oxygen is a result of the absence of sufficient mixing of the air by turbulence; instead, its composition is dominated by gas diffusion in the field of gravity. The region of 100 to 110 kilometres (60 to 70 miles) is one of transition, called the turbopause; above it, an element can exhibit its own natural vertical distribution, rather than that characteristic of the general atmosphere.

Extraterrestrial effects. Above a certain level the atmosphere is subject to ultraviolet radiation, to X-rays, and to solar particles. These cause the production of electrically charged particles—that is, ions and electrons from various kinds of atoms and molecules—in the ionosphere (*q.v.*), which extends from the mesosphere to the outermost limits of the atmosphere. But the charged particles are affected by the magnetic field of the Earth and consequently behave differently from the neutral particles in the air. In regions where the pressure is high enough, as in the mesosphere and the greater part of the thermosphere, ionospheric conditions are dominated by the preponderant neutral atmosphere. But, when the numerical ratio of charged particles to neutral particles is no longer negligible, the ionosphere is characterized by conditions in which account must be taken of the electric field connecting the positively and negatively charged particles. The region in which charged particles have energies greater than those corresponding to thermal velocities and move among the lines of force of the terrestrial magnetic field is the magnetosphere. It is a vast region primarily related to the interplanetary space in which the charged particles coming from the Sun are propagated. Another important influence is an auroral zone called the auroral oval (because of its form), characterized by the occurrence of polar auroras resulting from an influx of charged particles. Within the auroral oval the solar protons produce a particular type of ionization. Thus, there is also a geographic division of the atmosphere, resulting from the presence of the terrestrial magnetic field.

The region of the ionosphere

Composition of the atmosphere. The air near the Earth's surface has a well-defined chemical composition, consisting of molecular nitrogen, N_2 (78.1 percent by volume), molecular oxygen, O_2 (21 percent), argon (0.9 percent), and a small amount of carbon dioxide, CO_2 (0.03 percent). Also contained in the atmosphere (see Figure 2) are small, variable amounts of water vapour, H_2O, and trace quantities of methane, CH_4, nitrous oxide, N_2O, carbon monoxide, CO, hydrogen, H_2, and ozone, O_3, and of helium, neon, krypton, and xenon.

Where there is sufficient mixing by turbulence, the air consists essentially of molecular nitrogen and oxygen. Only at the altitude where molecular diffusion outweighs eddy diffusion does there begin the relative impoverishment of O_2 and N_2 molecules in comparison with atomic oxygen, O (Figure 2). The mass of an oxygen atom is 16 in relative units (*i.e.*, as compared to the mass of the atom of the isotope carbon-12, which is taken as the standard unit, a mass of 12), whereas that of molecular oxygen is 32 and that of molecular nitrogen 28. The mean mass of the air of the homosphere, 29, is gradually altered

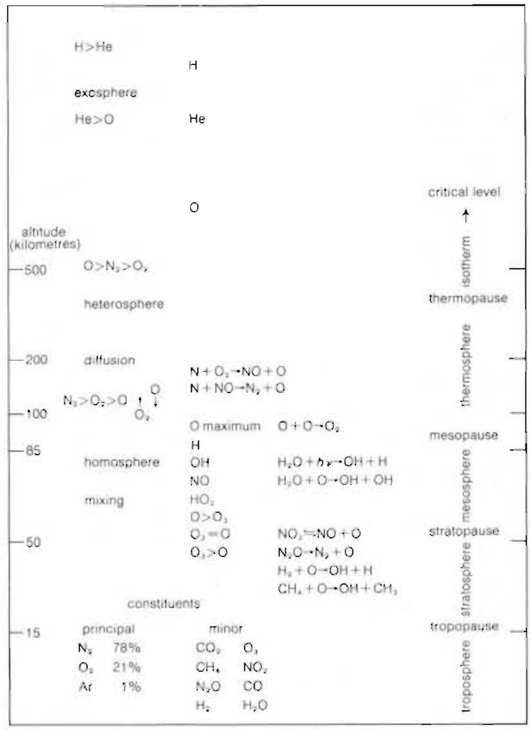


Figure 2: Distribution of principal and minor constituents of the homosphere, heterosphere, and exosphere.

above an altitude of about 100 kilometres (60 miles) to that resulting from an increasing proportion of atomic oxygen. In addition, the elements that are rare in the homosphere (for example, helium, $5/1{,}000{,}000$ of the air at sea level, and hydrogen, $1/1{,}000{,}000$ at sea level) become relatively more abundant at higher altitude. Because of their low masses (helium 4, hydrogen 1), the concentration of these atoms decreases much more slowly than that of oxygen and nitrogen in the heterosphere, where each element is independently subject to the field of gravity. Ascending into the heterosphere, the belts of molecular nitrogen and oxygen, of atomic oxygen, and of helium are encountered in that order. Finally, at its extreme limits, the atmosphere is composed of atoms of hydrogen.

Atomic mass and distribution

THE HOMOSPHERE

It is not difficult to describe the homosphere in terms of only the concentration of its principal elements, molecular nitrogen and oxygen; however, the phenomena that occur under the influence of solar and cosmic radiation must be taken into account as well. Although most of the molecules of nitrogen and oxygen are not affected by radiations originating outside the atmosphere, the absorption of various radiations does lead to certain ionization and dissociation effects.

Galactic cosmic radiation produces a permanent ionization in the whole of the stratosphere and in the mesosphere. In the homosphere, however, ionization is strictly a secondary phenomenon; dissociation under the influence of solar ultraviolet radiation is more important.

The effects of ultraviolet radiation. Molecular oxygen, O_2, is photodissociated under the influence of radiation of wavelengths less than 2400 Å (angstroms; the angstrom is a unit of length equal to 10^{-8} centimetre) into atoms of oxygen, O. This process is the foundation of all photochemical reactions in the homosphere. Production of atoms of oxygen, O, in the stratosphere and the mesosphere leads immediately to the production of molecules of ozone, O_3, as is shown in the equation

$$O_2 + O + M \rightarrow O_3 + M,$$

in which a triple collision between a molecule of oxygen O_2, an atom of oxygen O, and a third particle, M, which may be a molecule of oxygen or of nitrogen, results in formation of a molecule of ozone, O_3.

In the homosphere, the ozone, O_3, and molecular oxygen, O_2, molecules absorb all the solar ultraviolet radiation from 3000 to 1800 angstroms (ultraviolet ranges in wavelength from 100 to 3900 angstroms). Ozone absorbs primarily the radiation between 3000 and 2000 angstroms, and molecular oxygen that below 2000 angstroms. It follows that the absorption of ultraviolet radiation by ozone is the basis of the increase of temperature with altitude in the stratosphere; the ultraviolet heating is sufficient to develop a maximum temperature at the stratopause. Conversely, absorption by oxygen is very weak in the mesosphere, with the result that temperature decreases with altitude to the very low minimum found at the mesopause.

Dissociation of ozone. The presence of ozone in the stratosphere and mesosphere sets in motion a series of chemical processes. Under the influence of solar ultraviolet radiation of wavelengths below 3300 angstroms, ozone is photodissociated into molecules and atoms of oxygen in excited states. Production of an excited oxygen atom is important because it is a powerful oxidizer (*i.e.*, capable of producing chemical change by destroying normal molecules such as water vapour, methane, molecular hydrogen, and nitrous oxide).

When an excited oxygen atom is in the presence of, for example, stratospheric water vapour, H_2O, an immediate oxidation takes place, producing two hydroxyl radicals, OH (a radical is a group of atoms that occurs in the molecules of many compounds), expressed by the equation

$$O + H_2O \rightarrow OH + OH.$$

Formation of hydrogen compounds. With the dissociation of the water vapour, many compounds of hydrogen can be formed; for example, the hydroperoxy radical, HO_2, and hydrogen peroxide, H_2O_2. Furthermore, because the oxides of nitrogen NO and NO_2 are present, such compounds as nitric acid, HNO_3, can be formed. Analysis of these reactions shows that within the homosphere the existence of a mild degree of photodissociation of oxygen results in a remarkable series of reactions, which are important to the nature of the stratosphere and of the mesosphere. Accordingly, within the homosphere must be introduced the concept of the chemosphere, within which such chemical reactions occur.

Mechanisms of air mixing. One of the essential characteristics of the homosphere is that the turbulence of the air is always sufficient to maintain the same proportions of molecular nitrogen and oxygen in the mixture. Although molecular diffusion—that is, the tendency of gases of different molecular weights to separate—exists in the homosphere, it is always counterbalanced by eddy diffusion—the tendency to perfect mixture, in the case of the principal gases. But in the case, for example, of methane and of molecular hydrogen, which are oxidized as is water vapour, their removal from the air cannot be compensated by an influx of molecules from the troposphere. Molecular diffusion and turbulent diffusion are not sufficient to compensate for the loss resulting from oxidation. At the level of the stratopause, methane begins to disappear from the atmosphere. Water vapour is still present, however, because of its rapid recombination in the stratosphere and in the mesosphere.

The examples given illustrate the importance of the mechanisms of horizontal and vertical transport of the air in the stratosphere and in the mesophere. Analysis of the homosphere requires that the dynamic aspect be taken fully into account; further, the presence of helium and hydrogen in the heterosphere depends upon the conditions of transport in the homosphere. The helium produced in the Earth's crust enters the atmosphere and makes its way through the homosphere by the process of molecular diffusion until it reaches the thermosphere, where it plays an important role that will be discussed elsewhere in this article; atomic hydrogen is the product of the oxidation of hydrogen compounds, which occurs as determined by the properties of the homosphere.

THE HETEROSPHERE

At the beginning of the lower part of the thermosphere the change of composition and the increase in temperature are evident. Atomic oxygen becomes a factor in determining the mean molecular weight at an altitude of about 100 kilometres (60 miles). In addition to the atomic weight, 32, of the oxygen molecule, O_2, and 28, of the nitrogen molecule, N_2, the weight, 16, of atomic oxygen, O, must be considered. The decrease of the mean molecular weight essentially results from the photo-dissociation of the oxygen molecule under the influence of solar ultraviolet radiation of wavelengths below 1750 angstroms. Dissociation also occurs between 2000 and 2400 angstroms.

Oxygen dissociation in the lower thermosphere. An examination of thermosphere conditions reveals that equilibrium conditions cannot exist at each level between dissociation and reconstitution of the oxygen molecule. In other words, equality between the number of reactions that result in dissociation of the oxygen molecule and the number that result in recombination of the oxygen atoms is not possible under the conditions imposed by the altitude. At 100 kilometres the molecule can exist for more than ten days without being photo-dissociated, and accordingly its vertical distribution depends much more on the transport conditions than upon conditions of formation or destruction.

Variance in equilibrium conditions

There is, in fact, a departure from the conditions of photo-dissociation equilibrium and an approach to conditions of diffusion equilibrium. As a result, the following mechanisms are observed in the upper thermosphere: (1) when a molecule of oxygen is photo-dissociated by solar radiation, it yields two atoms of oxygen which do not recombine at the altitude at which they were produced, but are subject to a diffusion current descending to the altitude at which the atmosphere is dense enough to cause their disappearance; (2) the dissociated molecule is replaced by a molecule arriving from below, because O_2 survives long enough in the field of solar radiation to be transported upward by molecular diffusion.

In short, the dissociation state of the oxygen in the lower thermosphere is not determined by the equilibrium of photo-dissociation; instead, heavy O_2 molecules are carried upward, whereas the light atoms of oxygen must descend before they can recombine.

Effects of vertical transport. In the region in which the oxygen is subject to conditions of vertical transport, the transport processes also apply to other constituents of the atmosphere. At the altitude between 100 and 120 kilometres, eddy diffusion begins to lose its superiority over molecular diffusion. The transition results from the fact that the coefficient of molecular diffusion is inversely proportional to the density. Between sea level and 100 kilometres the density has diminished by a factor of at least 1,000,000,000; accordingly, molecular diffusion has increased in intensity by the same factor. Eddy diffusion having, on the other hand, gained no special advantage at 100 kilometres, it follows that molecular diffusion has become predominant at 120 kilometres. As a result, there is time for all of the atmospheric constituents that are not exceptionally reactive to become distributed in the gravity field according to their individual masses.

Atmospheric distribution by mass

Distribution of oxygen and nitrogen. Nitrogen, N_2, which is the most abundant molecule at 100 kilometres (of the order of 10^{13} molecules per cubic centimetre) and has a mass of 28, follows practically the vertical distribution of normal air, of which the mean molecular weight is 29. Molecular oxygen, O_2, with a mass of 32, decreases more rapidly with altitude than does molecular nitrogen (Figure 3).

Atomic oxygen, O, of mass 16, shows a relative increase with altitude in comparison with the oxygen and nitrogen molecules; it becomes more abundant than molecular oxygen, O_2, above 125 kilometres and more abundant than nitrogen, N_2, above 250 kilometres. As a result, the heterosphere is characterized by a thick layer of atomic oxygen extending for several hundreds of kilometres.

Distribution of helium and hydrogen. When subject to the diffusion process, helium and atomic hydrogen, whose proportions at 100 kilometres are of the order of $1/1,000,000$, can become the predominant elements because of their low masses (helium, 4; hydrogen, 1). It can be

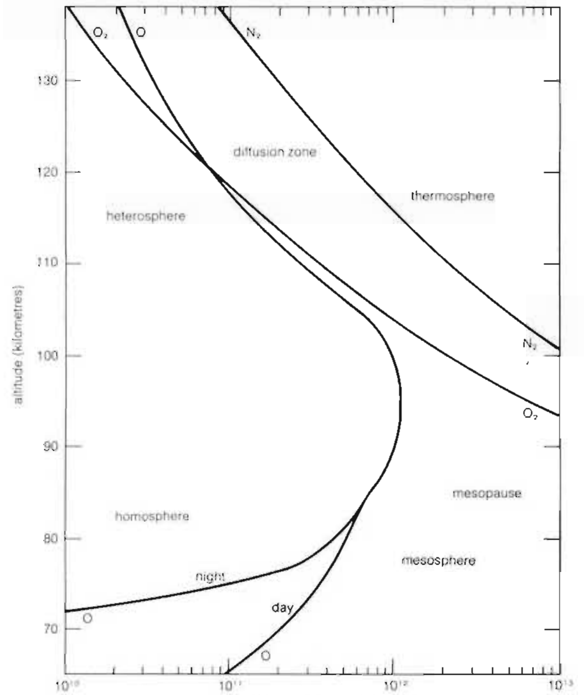
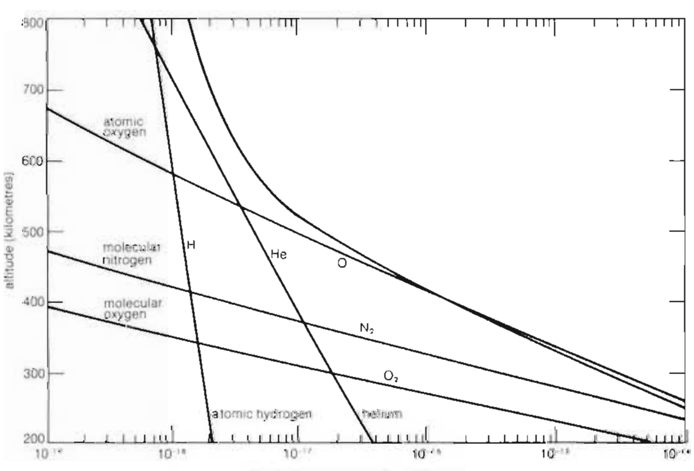


Figure 3: Concentration of oxygen and nitrogen in the atmosphere with respect to altitude.

shown that a belt of helium exists above the layer of atomic oxygen and also that atomic hydrogen forms the uppermost levels (corona) of the terrestrial atmosphere. The vertical distributions of atomic hydrogen, helium, atomic oxygen, molecular nitrogen, and molecular oxygen are presented in Figure 4, in which the density at a relatively low temperature of 600° K is given for each constituent.

Variations in atmospheric densities. Review of the data obtained from the drag on artificial satellites makes it evident that the density of the heterosphere is subject to considerable variations (Figure 5). Whereas at the maximum of solar activity that occurred in 1958 the density was extraordinarily high, it came down to a very low value during the 1964 minimum of solar activity. There is, thus, a close correlation with solar activity. It is the ultraviolet radiation, which penetrates into the heterosphere and is absorbed there, that exhibits variations closely linked to the Sun's activity. To the 11-year cycle of solar activity must be added the variation corresponding to the rotation of the Sun. Finally, it must be emphasized that above 200 kilometres there is a remarkable variation of density between day and night.

The rate of ultraviolet radiation



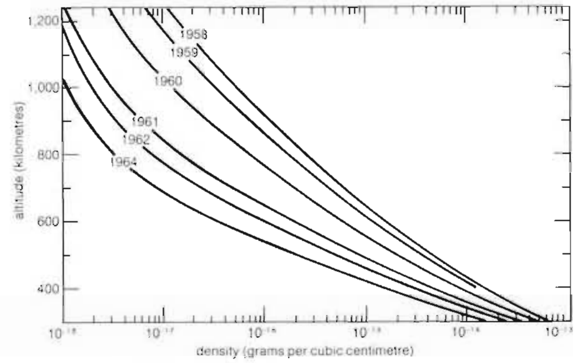


Figure 5: Relationship between mean atmospheric density and solar activity in the heterosphere.

Temperature fluctuations. To interpret the density variations in the thermosphere, it is important to consider that the thermospheric gas is subjected to strong variations of temperature. The thermal balance is evaluated using the assumption that an ultraviolet heating effect and a rapid recooling work against each other. Thermal conditions in the heterosphere evidently differ from those of the homosphere. In the lowest region of the homosphere, the troposphere, which is in contact with the ground, the source of heating is essentially the ground, which has absorbed the Sun's rays, and convection is an adequate means of transporting the heat. In the stratosphere and mesosphere, convection and infrared radiation from one direction meet ultraviolet heating from the other. But in the thermosphere the third means of heat transport—conduction—decisively outweighs the other means, radiation and convection. For that reason the thermal balance is such that very different conditions appear by day and by night.

During the day, the cooling by conduction is compensated by ultraviolet heating, whereas during the night the heat loss by vertical conduction is not compensated by horizontal transport. As a result, there is a very large daily variation of the temperature of the thermopause, which is manifest to the observer as a variation of the atmospheric density.

Cycles of solar activity. Furthermore, the fact that periods of 27 days can be detected in the variations of atmospheric density must be interpreted in terms of the rotation of the Sun. Ultraviolet emission does take place when the Sun is perfectly calm over the whole of its disk; but the brilliant regions visible as sunspots emit a greater ultraviolet radiation. An increase of solar activity in the hemisphere of the Sun that is visible from the Earth is an index of increased ultraviolet heating. Accordingly, the thermal balance in the thermosphere varies according to the amount of ultraviolet heating available to counter the conductive cooling.

Taking account of the variation in ultraviolet radiation during a cycle of solar activity, it is possible to predict the long-term variations of the thermospheric temperature. In Figure 6 the maximum variation of daytime temperature is shown; the mean temperature varies from a low of 750° K, with minimum solar activity, to a high of about 2,000° K.

Because the composition of the heterosphere varies as a function of temperature, the vertical distribution of the atmospheric constituents is variable (Figure 7). The proportion of helium is greatest at the highest temperatures and the proportion of hydrogen is greatest when the temperature is lowest.

The escape of helium and hydrogen. It must be remembered that the aeronomic problems regarding helium and atomic hydrogen are essentially different from those associated with the other elements. The continual introduction of helium into the atmosphere—at an average rate of about 1,000,000 atoms per square centimetre per second (as a result of its formation in the Earth's crust by the disintegration of thorium and uranium)—would lead in the course of some millions of years to a doubling of the total quantity of helium existing in the atmosphere.

Sources of airborne helium

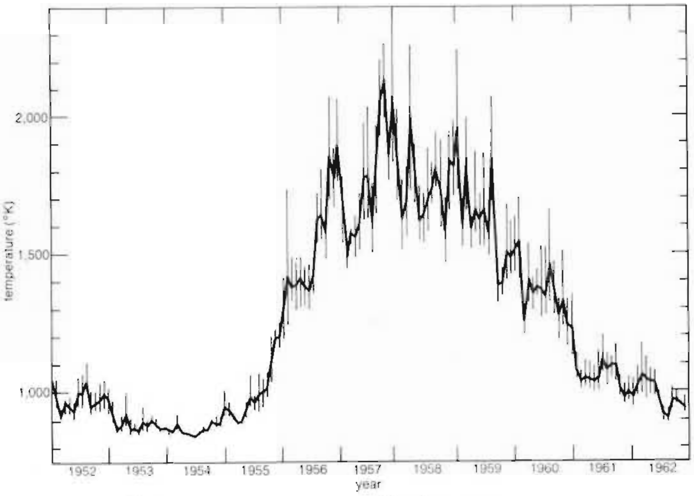


Figure 6: Mean temperatures at the thermopause for 27-day periods during 1952–1962 (see text).

Since the formation of the Earth about 4,600,000,000 years ago, an amount of helium equal to that introduced from the above sources must have escaped to avoid accumulation. Such escape can occur only at the top of the atmosphere by a transport process taking place in the homosphere and the heterosphere. The equations of diffusion reveal that the escape of 1,000,000 atoms of helium per square centimetre per second is perfectly possible, thus maintaining a helium distribution close to the equilibrium of diffusion in the heterosphere.

In the case of hydrogen, of which some 100,000,000 atoms escape, the limit of what diffusion will support is reached. For this reason, the vertical distribution of hydrogen is not that of a diffusion equilibrium but depends on a transport state in which the number of atoms moving toward the base of the heterosphere is equal to the number leaving it. It can be concluded that there are more atoms of hydrogen in the thermosphere at low temperature than at high temperature. The role of hydrogen in the thermosphere is therefore most evident at the minimum of solar activity. In the last analysis, the total mass of hydrogen atoms passing from the terrestrial atmosphere into interplanetary space is several hundred grams per second.

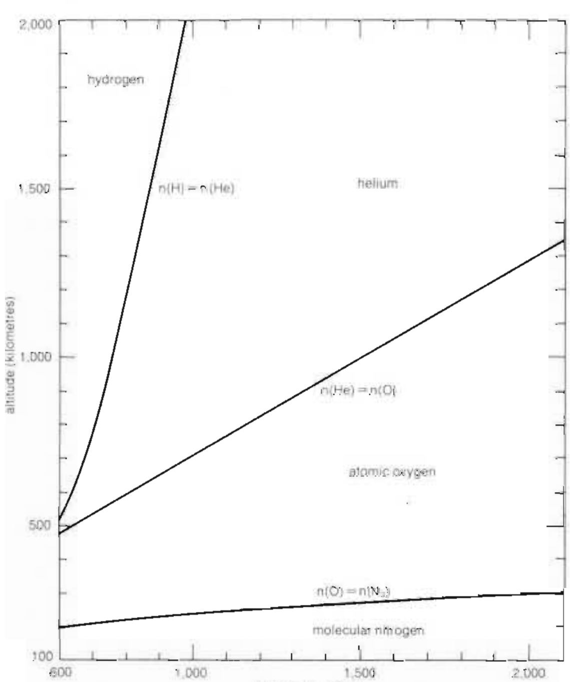


Figure 7: Atmospheric zones of common gases as a function of temperature and altitude.

It is possible to go further into the subject of the heterosphere with respect to perturbations such as the magnetic storms. It is clear that the contribution of energy due to certain particles can cause a general heating only if the energy is capable of penetrating across the lines of force of the terrestrial magnetic field. The heating can be local, as in the auroral oval, where there is direct precipitation of particles. It has been shown that the density of the thermospheric gas varies with the geomagnetic index, which can be explained by a variation of temperature. Such processes, however, require complicated explanations.

Ionization and solar activity

Ionization mechanisms. Because the aeronomic conditions of the ionization state depend simultaneously on the vertical distribution of the neutral elements nitrogen and oxygen and on the spectral distribution and the absorption of the solar radiation, it may be readily concluded that the amount of photo-ionization is closely related to the diurnal, seasonal, periodic, and 11-year variations in solar activity. In addition, account must be taken of the reactions occurring between the ions and the neutral particles and also of the processes that result in the disappearance of electrons and ions.

When an ionizing radiation causes the ejection of an electron from an atom or a particular molecule, a photoelectron appears, which possesses kinetic energy corresponding to the difference between the solar radiation energy and the energy required for the ionization. Such a photoelectron can give up part of its energy to ionize another atom if its energy is sufficiently high, as is the case with strongly energized electrons produced by X-rays. In general, any photoelectron loses part of its energy in exciting the most plentiful atoms and molecules; however, in the upper ionosphere, photoelectrons colliding particularly with the other electrons cause a general heating of the electronic gas.

The nitrogen ion. A photo-ion (atomic or molecular ion produced by photo-ionization) can collide with an electron, and it is possible that the two will reconstitute the neutral element. But it is more likely that the photo-ion will collide with an atom or molecule of another kind, exchange its charge, and return to the neutral state; for example, the nitrogen molecule—the element most plentiful below 200 kilometres—should furnish the most abundant ion. In reality, the photo-ion of molecular nitrogen very quickly comes into contact with a molecule of oxygen or with an atom of oxygen. The first case involves a transfer of charge; that is to say, the peripheral electron of the oxygen molecule, O_2, is transferred to the nitrogen ion, N_2^+, which reconstitutes the normal molecule, N_2, and forms the molecular ion, O_2^+: $N_2^+ + O_2 \rightarrow N_2 + O_2^+$.

The second case corresponds to a change of the atom and of the ion, in which the molecular nitrogen ion, N_2^+, colliding with the oxygen atom, O, dissociates to form an ion of nitric oxide, NO^+, and an atom of nitrogen, N: $N_2^+ + O \rightarrow NO^+ + N$. Thus, the second case results in an NO^+ ion that has been produced directly by the action of solar radiation. A molecule of nitric oxide appears in the heterosphere by way of the formation of the ion. In any case, the formation of the NO molecule in the thermosphere results in its descent toward the mesophere, where it is subject to the dissociation effect in that region.

Equilibrium exchange of atomic charge

Atomic-hydrogen ions. Although the hydrogen atom is photo-ionized by solar radiation, its ionization state is determined primarily by its reaction with atomic oxygen. There is an equilibrium of exchange of charges closely related to the neutral or ionized condition of the atmosphere. This equilibrium is written:

$$O^+ + H \rightleftharpoons H^+ + O,$$

in which an atomic oxygen ion, O^+, and a neutral atom of hydrogen, H, react to produce a neutral atom of oxygen, O, and a hydrogen ion, H^+, and vice versa. This equilibrium results from the fact that the two atoms have nearly equal ionization energies, and, if the concentrations are great enough, they will pass readily from one state to the other in accordance with the proportions of the two constituents, H and O, present in the neutral state. But ionization equilibrium conditions between atomic oxygen and hydrogen cannot be maintained at the highest altitudes. The mechanism of exchange of charge depends on the frequency of the collisions of the ions with neutral atoms, whose concentration decreases with increasing altitude. Consequently, when the exchanges are infrequent the phenomenon of diffusion of the ions in the gravity field appears. The hydrogen ion, H^+, which is only $\frac{1}{16}$ as heavy as the oxygen ion, O^+, will have a tendency to distribute itself along a line of force of the magnetic field in accord with its particular weight and to escape reaction with the neutral oxygen atom. As a result, atomic-hydrogen ions become much more numerous than atomic-oxygen ions.

THE EXOSPHERE

When it is stated that the temperature is about 273° K at sea level or at 50 kilometres altitude or that it is about 1,500° K at 500 kilometres, the same physical basis of determination is used, and it is incorrect to suppose that a change of concept is involved. The thermometer currently used in meteorology is not sufficiently sensitive to measure the exact temperature of the air at high altitudes. Even at sea level, many precautions must be taken in order to obtain an exact measurement of the air temperature.

Division of molecular energies

The temperature of gases. The kinetic theory of gases states that the molecules are characterized by very rapid motions, with a distribution of velocities such that a mean velocity corresponding to a specific temperature can be determined. When the collisions are very numerous, there is an equal division of the energy among the various kinds of molecules, which defines a unique temperature. To state that the temperature is of the order of 273°K at a certain altitude means that the kinetic energies of the molecules are identical, because there are enough collisions to assure equal division of the energy. Consequently, perfect thermometers (which could attain equilibrium immediately) would give the same indication. No such thermometers can be made, however, and, accordingly, indirect procedures are used, based on the determination of pressure and on the law of perfect gases.

Reactions between particles. When the heterosphere is analyzed in terms of its physical constitution, in which electrons, ions, and neutral atoms are simultaneously involved, it is necessary to go into the details of various types of reactions and to determine their exact nature. First of all, it must be remembered that the interaction between two electrons is much more marked than that occurring between an ion and an electron. Also, the interaction between an electron and a neutral atom is much weaker than that between an ion and an electron. All these interactions can take various forms, ranging from elastic collisions to exchanges of charge or of energy. It follows that the electrons can have their own temperature when there are so many electrons relative to neutral particles that the interaction between electrons predominates over other interactions. Moreover, when the time required for exchanges of energy in interactions between ions and neutral particles is relatively long, it is possible for the ions and the neutral atoms to be maintained at different temperatures.

Collision frequency and temperature differences

Distribution of energy exchange. The problem of the temperature of a neutral gas, ions, or electrons can be described as follows: a photoelectron ejected, under the action of solar radiation, from a neutral atom of oxygen possesses a certain kinetic energy distinctly greater than that of the electrons already present at the same altitude. In the lower thermosphere (below 120 kilometres), the photoelectron will collide with the relatively numerous neutral molecules, rapidly lose its excess energy to them, and reach the temperature of the neutral gas. In the middle thermosphere (above 150 kilometres), the photoelectron will come into contact primarily with the electrons already present, and its excess energy will serve initially to increase their total energy; the result will be an increase of the temperature of the electrons as a group relative to the temperature of the ions and to that of the neutral atoms. In the upper strata, high-energy photoelectrons will come initially into contact with the other electrons that undergo collisions with the ions. Again, the photoelectrons will cause a rise in the electron temperature and indirectly in that of the ions, but, just as in the neutral atmosphere, it must be taken into account that heat conduction exists within an electronic gas exhibiting a temperature gradient. Consequently, there will be a tendency to isothermy at the highest altitudes by reason of the rapid heat transport. Thus, it is evident how greatly the physical state of the neutral atoms, the ions, and the electrons in the upper atmosphere differs from that created in the laboratory.

Atom velocities at the critical level

The gradation of particle collisions. Beginning at the critical level, temperature loses its ordinary meaning, because the particles follow their free-space trajectory and practically do not undergo collisions. Because of this it must be kept in mind that, although for a given temperature the atoms have the same energy, they do not have the same velocity; the latter depends on their mass. The light atoms move with greater velocities than the heavy ones; for a given energy, hydrogen has double the velocity of helium, which in turn has twice the velocity of atomic oxygen. Consequently, in the exosphere a light atom such as hydrogen can escape from the terrestrial atmosphere because of its velocity, four times that of atomic oxygen, which remains permanently subject to the gravity field.

Assuming that all the elementary physical processes can operate without restriction in the heterosphere, it is then clear that the changes of structure and composition can be explained within the usual framework of a hydrostatic equilibrium. The collisions are always sufficiently numerous to provide a complete distribution of particle velocities as defined in the kinetic theory of gases. This is what is called a Maxwell distribution, leading to a mean kinetic energy of the particles and a clear definition of the temperature. With increasing altitude above the thermopause, however, the mean free path of the atoms gradually increases. Beyond a certain altitude this mean free path is so great that collisions between the atoms are too few to maintain the uniformity of the atmospheric gas. The particles—atoms, ions, electrons—can traverse long distances under the influence only of the field of gravity in the case of the neutral particles, and only of gravity and the magnetic and electric fields in the case of the charged particles.

Thus, the heterosphere, in which the collisions must be taken into account, is separated from the exosphere, where the collisions can be neglected, by a critical zone in which transition occurs from the point, at its base, where the number of collisions is significant to the zone above it, where the number of collisions is negligible.

Determination of the critical zone. The most satisfactory means of determining the critical zone is to search for a level at which the aeronomic conditions can be defined for each element, neutral or ionized. It is possible, for example, to study the mean free path of each element. To specify the probability that, for example, in one out of two instances a particle arriving at a particular level will be able to escape without undergoing a collision is also to specify the mean free path characterizing a critical level corresponding to the base of the exosphere.

Neutral and ionic exosphere differences

Computation shows that, for the neutral atoms, the critical level for temperatures ranging from 750° to 2,000° K is located at altitudes ranging from 400 to 800 kilometres (250 to 500 miles), respectively. In other words, the neutral exosphere, considered as the region in which the collisions between neutral atoms are so infrequent as to permit the escape of atoms whose velocity is sufficient (11 kilometres per second), begins at 400 kilometres with minimum solar activity and can reach 800 kilometres with maximum solar activity. The ionic exosphere begins at an altitude 2,000 or 3,000 kilometres (1,000 or 2,000 miles) higher than that of the neutral exosphere, because collisions between charged particles are always more numerous. The ionic exosphere differs in another respect from the neutral ionosphere: although the gravity field influences the trajectory of the charged particles as it does in the neutral exosphere, the charged particles also are controlled by the geomagnetic field and by the electrostatic field between heavy ions and electrons.

Particle trajectories in the exosphere. In any case, a given volume of the exosphere can contain a certain number of particles having a priori the following trajectories (see Figure 8): (1) particles coming from the critical level and returning there; (2) particles coming neither from the critical level nor from interplanetary space; (3) particles coming from the critical level and proceeding to interplanetary space or vice versa; (4) particles coming from and returning to interplanetary space. These four groups represent all the possibilities arising from a Maxwell distribution of velocities; it can accordingly be said that at any altitude the density, ρ, of a given element normalized to unity is composed of four densities:

$$\rho = \rho_{I} + \rho_{II} + \rho_{III} + \rho_{IV} = 1.$$

Particles with ballistic trajectories. The first group (see Figure 8) consists of particles with ballistic trajectories: those that leave the critical level at a certain angle with a certain velocity and return to it after having attained their limiting altitudes in the exosphere. The kinetic energy at departure from the critical level is less than that required for escape from terrestrial attraction. This is the case of the neutral particle the trajectory of which is regulated by the gravity field; but a charged particle is constrained to follow the lines of force of the geomagnetic field. A completely closed line of force does not permit the escape of an ion or an electron, whatever its velocity.

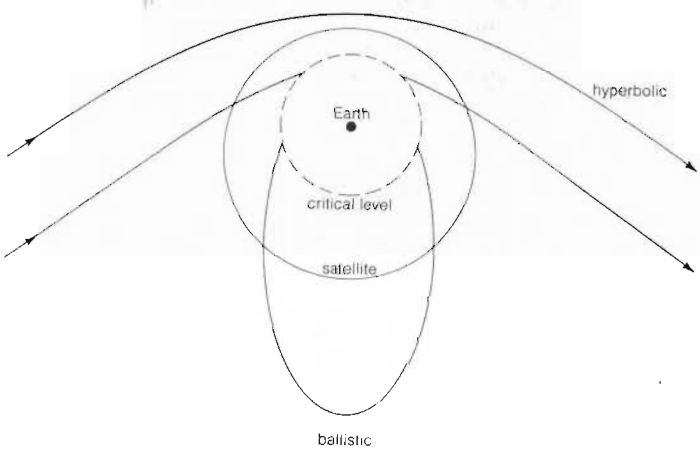


Figure 8: Trajectories of particles in the exosphere. In the vicinity of the critical level, 400 to 800 kilometres above the Earth, ballistic trajectories are most numerous, whereas at much greater distances, hyperbolic trajectories are evident.

Significant and insignificant exosphere constituents

Trapped particles. The second group of particles (Figure 8), not making contact with the critical level or with the exterior, corresponds to trapped particles. These are satellite neutral atoms or ions, subject to backward and forward movements along a line of force of the magnetic field. It can be shown that the conditions to be satisfied in order to maintain thermal particles in elliptic orbits are difficult to achieve. In general, the presence of satellite or entrapped particles in the exospheric distribution can be disregarded; the density ρ_{II} of exospheric particles following elliptical orbits in the presence of particles of density ρ_{I} following ballistic orbits is not significant.

Particles with escape velocities. The third group of particles must be subdivided according to whether they leave the critical level and escape to interplanetary space or arrive at the critical level from interplanetary space. In general, it is not necessary to consider particles with adequate velocities arriving from interplanetary space; the significant component of the third group consists of atoms at the critical level capable of attaining velocities above the escape velocity of 11 kilometres per second. This is the case with the hydrogen atom and sometimes the helium atom when the temperature of the heterosphere is sufficiently high. In the case of the ionic exosphere, escape is possible only where the particles follow open lines of force of the geomagnetic field into free space. In general, the escape of charged particles from the ionic exosphere is neglected, because a line of force of the geomagnetic field joins one hemisphere to the other except near the poles. Only through the channel of the polar cap, where the lines of force of the geomagnetic field are open toward the nocturnal magnetosphere, do charged particles have the possibility of escaping. Indeed, such is found to be the case for ions of hydrogen and helium, which escape easily from the polar ionic exosphere.

Extraterrestrial particles. Finally, it is possible to conceive of particles moving at high velocities that arrive from the exterior of the exosphere and return without touching the critical level. These would comprise the fourth group. Such particles are naturally neglected in the exospheric population, as they never attain sufficient concentration.

To sum up, the population of the neutral or the ionic exosphere is composed almost exclusively of particles coming from the critical level. In fact, the ballistic trajectories represent the practical conditions of vertical distribution for such elements as atomic oxygen. The concentration of neutral helium differs according to whether the temperature is high or low. In the case of neutral hydrogen, the trajectories representing permanent escape of this atom from the terrestrial atmosphere into interplanetary space are significant compared with the ballistic trajectories. As for the ions, it may equally be said that it is the elements that come to the critical level that constitute the population of the ionic exosphere. The oxygen ion, which is too heavy, does not escape, whereas the ions of hydrogen and helium escape into the polar exosphere. Thus, the atmosphere of the Earth, terminating in its neutral or ionic exosphere, is populated at its extreme limits by neutral and ionized atoms of hydrogen.

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(M.N.)

Atmosphere, Development of

The development of the Earth's atmosphere is a subject on which there is little direct evidence. Conjecture is limited only by geological consequences attributable to atmospheric evolution and by available theory. Among the important questions to which answers are sought are the origin of the gaseous components, the nature of the initial and early atmosphere, and the kinds of compositional changes that subsequently occurred in response to additions of some gases and losses of others. The answers to such questions must be sought within diverse areas of knowledge, because atmospheric evolution is related to