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Temperatures of the Earth's Upper Atmosphere

Aeronomers have discovered distinct temperatures for the different components of the ionospheric plasma.

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The temperature of the earth's upper atmosphere is one of the most revealing properties of the earth's near environment. Not only does it vary widely with time and location but it also reacts strongly to changes in solar activity. The variation of temperature with altitude and with time reflects directly the different energy sources which in large measure govern the dynamic behavior of the upper atmosphere. The temperature also controls the rate of change of density with altitude through the requirement of hydrostatic balance. In hydrostatic balance the atmospheric pressure at any height equals the total weight of the overlying gas, a condition which requires that the pressure and density of the gas decrease exponentially at a rate inversely proportional to the temperature (*1*).

Thus, if the altitude profile of the temperature is known, one can calculate the altitude profiles of pressure and density, provided the mean molecular weight of the gas is also known. This proviso is necessary because the rate of decrease of pressure and density is proportional to the mean molecular weight. Since heavy gases, such as argon and carbon dioxide, are more tightly bound by the earth's gravitational field, they tend to concentrate at low altitudes, while the density of light constituents, such as hydrogen and helium, decreases very slowly with height. At altitudes below about 110 kilometers, however, this tendency toward gravitational separation of the constituents is fully counteracted by turbulent mixing processes, so the mean molecular weight of the atmosphere varies very little from its sea-level value of 29 atomic mass units (amu). At higher altitudes there is little mixing. The heavy constituents become progressively more rare, and the dominant atmospheric constituent changes, with increasing altitude, from molecular nitrogen (28 amu), to atomic oxygen (16 amu), to helium (4 amu), and, at very high altitudes, to atomic hydrogen (1 amu).

The Upper Atmosphere

Different people have different ideas about where the upper atmosphere begins, depending on their points of view. In our discussion we take, as the base of the upper atmosphere, the level where the mean molecular weight begins to decrease.

The most abundant gases in the upper atmosphere are molecular nitrogen and molecular oxygen, atomic oxygen, and, at high altitudes, helium and atomic hydrogen (see Fig. 1). The total concentrations are very low—less than 10^{12} molecules per cubic centimeter at 120 kilometers and 10^6 per cubic centimeter at 1000 kilometers, compared with more than 10^{19} per cubic centimeter at the ground. At these low densities collisions between molecules are relatively infrequent, permitting chemically active species such as atomic oxygen, free ions and electrons, and metastable excited atoms and molecules to exist in appreciable numbers. Since it is almost impossible to attain comparable conditions of low density and large volume on the ground, the upper atmosphere is, for many, a laboratory for the study of species and processes which are difficult to produce and to detect in ground-based laboratories.

The gas densities in the upper atmosphere are low, but the temperatures are generally high. The mean temperature of the neutral gas increases from 350°K at 120 kilometers to about 1000°K at 250 kilometers; above that altitude it becomes very nearly constant. At the low existing densities, however, these high temperatures do not mean that there is a large source of energy in the upper atmosphere; the height-integrated heating rate is only about 2 ergs cm^{-2} sec^{-1}, about one-millionth of the rate at which the ground is heated by visible radiation from the sun. Instead, the high temperatures in the upper atmosphere reflect the inefficiency of mechanisms for removing heat from this region.

In the lower atmosphere heat is transported relatively efficiently from one altitude to another by turbulence and convection, but in the upper atmosphere turbulence is inhibited by viscosity and the positive temperature gradient renders the atmosphere stable against convection. In the lower atmosphere heat is also transported efficiently by infrared radiation emitted by water vapor, carbon dioxide, and ozone. In the upper atmosphere, on the other hand, infrared emission is relatively unimportant. The molecules cannot radiate energy any faster than they receive energy in collisions with other molecules. At very low densities such energizing collisions are rare.

As a result, the heat which is de-

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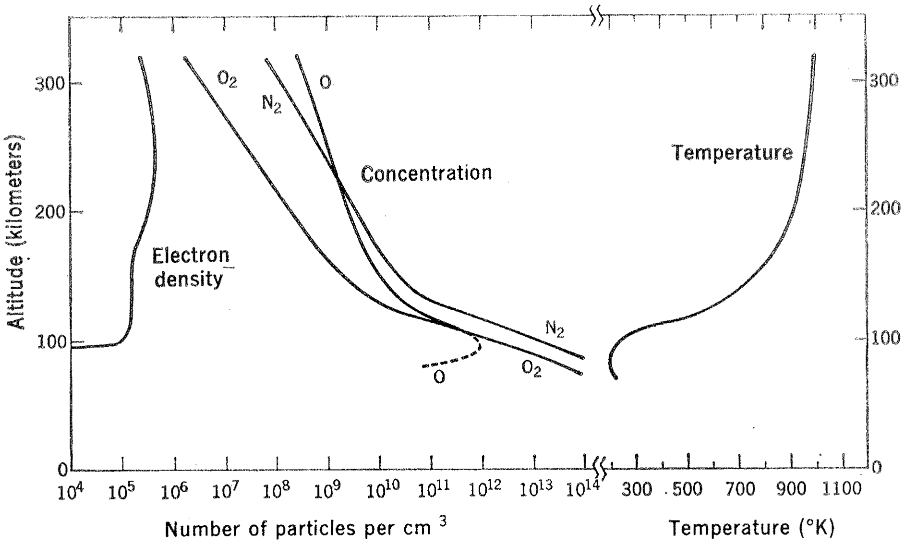


Fig. 1. Atmospheric densities based upon measurements made on NASA rockets 18.01, 18.05, and 14.285 and the corresponding temperature profile derived from the density profile by means of the hydrostatic balance relationship.

posited by solar radiation at altitudes of about 200 kilometers has to be removed from the upper atmosphere by thermal conduction (*2, 3*). The heat must be conducted all the way down to about 100 kilometers, where infrared emission can remove it. This is the reason for the substantial temperature gradient between 100 and 200 kilometers and for the high temperatures in the upper atmosphere.

Formation of the Ionosphere

The small upper-atmosphere heat source is provided by the absorption of extremely short wavelength ultraviolet radiation ($\lambda < 1000$ angstroms) from the sun. The absorption of this radiation results in photoionization, in which an electron is ejected from an atmospheric molecule. The rate of photoionization and consequent heating has a maximum at altitudes near 200 kilometers, falling off at high altitudes because of the decrease in the density of atmospheric molecules which can be ionized, and decreasing at low altitudes because of the attenuation of the ionizing radiation as it penetrates the atmosphere (*4*).

The free ions and electrons produced by photoionization undergo a number of exothermic chemical reactions (*5*), which contribute further to the heating of the upper atmosphere, before they recombine to form neutral atoms. The altitude profiles of electron and ion density in the ionosphere are determined by the balance between photoionization and recombination, the effects of diffusion of ions from one altitude to another being readily apparent at higher altitudes (*6*). Typically, the ion and electron densities exhibit a maximum of a little less than 10^6 per cubic centimeter at altitudes of about 300 kilometers and decrease to values below 10^4 per cubic centimeter at altitudes below 100 and above 1000 kilometers. The detailed behavior of the ionosphere depends also on the density and composition of the neutral gas and on the temperature. How, then, can the temperature of the upper atmosphere be determined?

Measurement of Temperature

It is not enough simply to mount a thermometer on a rocket or satellite. At high altitudes the density and heat capacity are so low that the air, in spite of its high temperature, would not feel hot to an astronaut who put his hand out the window of a satellite at an altitude of, say, 300 kilometers. The hand, the thermometer, and the satellite would all be heated by electromagnetic radiation from the sun and from the ground, not by the surrounding atmosphere, and they would be cooled by radiation also. Their temperatures would reflect the balance between absorption and emission of radiation and would generally be much lower than the gas temperature. The temperature of the atmospheric gas is a measure of the kinetic energy of the gas molecules, and the high gas temperatures mean that the molecules have comparatively high kinetic energies. In the upper atmosphere, however, the molecules are too few in number to transfer a significant amount of energy to a massive solid object such as a thermometer or a hand.

Before rocket and satellite exploration of the upper atmosphere had begun in earnest, a number of indirect arguments had already provided fairly convincing evidence that the temperature was high (*2, 7*). A particularly ingenious argument based on the abundance of helium was used by Spitzer in 1949 (*2*). Helium is produced continuously in the crust of the earth as a result of the decay of radioactive minerals, and the rate at which it is added to the atmosphere can be estimated from the concentrations of these minerals. It is clear that helium has not been accumulating in the atmosphere throughout geologic time, and there must therefore be a removal process.

At altitudes above about 500 kilometers there exists a region, known as the exosphere, where collisions are so infrequent that gas molecules can escape from the earth's gravitational field out into space, provided the velocity associated with their random thermal motion is sufficiently large. Spitzer assumed that helium escapes in this manner from the top of the atmosphere as fast as it is added at the bottom. Since the rate of this "evaporation" of the atmosphere into space is very sensitive to temperature, he was able to estimate the temperature of the exosphere. His value was 1500°K, surprisingly accurate in the light of recent evidence that processes other than "evaporation" contribute to the escape of helium from the atmosphere (*8*).

Another ground-based method of determining upper-atmosphere temperatures is provided by high-resolution spectroscopic measurements of the optical emission of the aurora and airglow (*9*). The Doppler shift of line emission from moving atoms gives information on the thermal motions of the atoms and hence the kinetic temperature. Alternatively, the rotational structure of the band emission from atmospheric molecules provides information on the rotational temperature, which is in most cases the same as the kinetic temperature. Unfortunately, it is not easy to determine the altitude from which the observed radiation is coming. Consequently, spectroscopic measurements have not proved very useful in determining temperature profiles in the upper atmosphere (*10*).

New Knowledge from Space Experiments

The source of most of our information on upper-atmosphere temperatures is a rather unlikely one. It is the orbital decay of artificial satellites. As a satellite moves through the atmosphere it is slowed by friction at a rate proportional to atmospheric density. The cumulative effect of this drag over a number of orbits leads to changes in the orbital period of the satellite which can be detected in tracking data (*11*). Since the launching of Sputnik I in 1957, observation of the decay of satellite orbits has provided a large body of data on upper-atmosphere density (*12*). Because the density profile depends upon the temperature through the requirement of hydrostatic balance, it has been possible to derive atmospheric temperatures from these densities.

Results from these satellite drag studies have shown conclusively that the temperature is high and have shown, in addition, that it is quite variable. At times of high solar activity, when sunspots are numerous and flares and eruptions on the surface of the sun are frequent, the daytime temperature in the upper atmosphere may be as high as 1700°K. At times of low solar activity this temperature may be as low as 900°K (see Fig. 2). These temperature variations result largely from variations in the extreme ultraviolet radiation from the sun, which, as noted above, is the principal heat source for the upper atmosphere.

In addition to this solar activity effect, there is a large diurnal variation in upper-atmosphere temperature, with afternoon temperatures exceeding predawn temperatures by a factor of 1.3. An equivalent excursion in the temperature at the ground would be 100°K! The sensitive response of temperature to diurnal change in the solar heat input is a consequence of the low density and correspondingly low heat capacity of the upper atmosphere.

The temperatures derived from satellite drag densities are subject to some uncertainty because the composition of the upper atmosphere is poorly known (*13*), and the variation of density with altitude depends on mean molecular weight as well as on temperature. Additional uncertainties result from the paucity of satellite drag data at altitudes below 200 kilometers, where densities are high and satellite lifetimes are short (*14*).

Rocket Measurements

Because of these limitations of satellite drag data, rocket measurements made with mass spectrometers have been particularly valuable. These experiments determine the density profiles of individual neutral species at altitudes ranging between 130 and 300 or more kilometers. The variation of the temperature with altitude can be deduced unambiguously from the density profile of a single constituent by means of the hydrostatic balance relationship. An alternative approach is to measure the thermal velocity distribution of the gas molecules directly, by means of a mass spectrometer with a velocity selector.

However, rocket measurements can be made from only a few locations in the world, and thus give very limited coverage in space as well as in time. Accordingly, they are not well suited to unraveling the complex variations of upper-atmosphere temperature. Perhaps the largest body of rocket mass-spectrometer data has been obtained in a series of experiments, involving use of an instrument known as the thermosphere probe, conducted jointly by scientists at NASA's Goddard Space Flight Center and the Space Physics Research Laboratory of the University of Michigan. Since 1962, the concentration and temperature of molecular nitrogen in the upper atmosphere have been measured in 22 successful flights at various times of day and under varying conditions of solar activity, from launch sites at Fort Churchill, Manitoba, at Wallops Island, Virginia, and at Vega Baja, Puerto Rico (*15*).

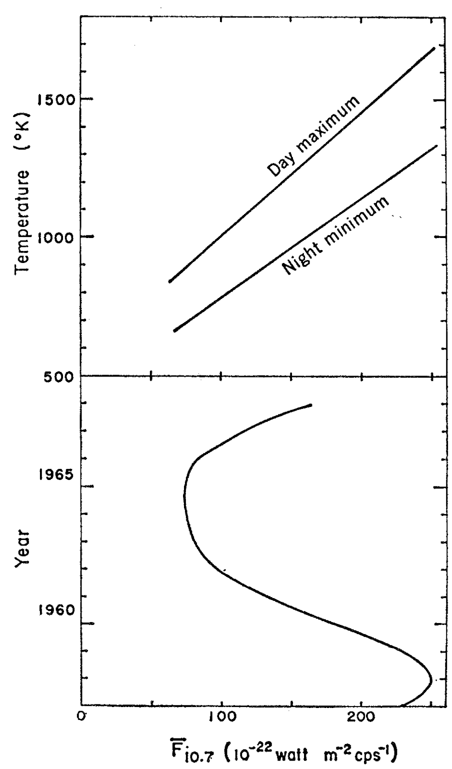


Fig. 2. Daytime maximum and nighttime minimum values for the temperature at the top of the atmosphere as a function of solar activity. As an index of solar activity we use the 10.7-centimeter solar radio flux ($\overline{F}_{10.7}$) averaged over several solar rotations. The variation of $\overline{F}_{10.7}$ over the last 10 years is also shown (see *12, 33*).

These experiments are the outgrowth of a program of direct measurement of ionospheric properties started at the University of Michigan in the late 1940's with V-2 rockets (*16*). In the earliest experiments, in 1946 and 1947, electrostatic probe techniques which had proved valuable in the study of laboratory plasmas were used. The electrostatic probe measures the current flowing to an electrode immersed in the plasma as a function of the voltage applied to the electrode. When the applied voltage is such as to attract electrons from the plasma, the current provides a measure of the electron density. By sweeping the probe through a range of retarding potentials, information can be obtained on the electron velocity distribution and thus on the electron temperature (*17*).

Temperature of the Ionospheric Electrons

These initial, exploratory experiments were not sufficiently sensitive to provide quantitative data. However, the program was resumed in 1958 with greatly improved instrumentation, and electron-temperature and density data were obtained from a rocket flight at Fort Churchill (*18*). The results were rather surprising. They suggested that the electron temperature exceeded the neutral-gas temperature, but the combined uncertainties in the data and in the neutral-gas temperature were too large to allow definite conclusions to be drawn.

By 1960 considerably more was known about the temperature of the neutral atmosphere, and measurements obtained in three rocket flights showed clearly that the probes were measuring electron temperatures considerably above the neutral-gas temperature both at Fort Churchill and at Wallops Island (*19*). These results stimulated several theoretical studies which, while differing in details, all indicated that the electron temperature should be higher than the neutral-gas temperature over a broad altitude region centered at 200 kilometers, but should approach the neutral-gas temperature at higher and lower altitudes (*20, 21*).

Because of their small mass, electrons lose very little kinetic energy when they collide with a heavy ion or neutral molecule. They simply bounce back like a tennis ball off a wall. On the other hand, energy is exchanged readily in collisions between electrons. As a result, the distribution of thermal velocities of the electrons is very close to the Maxwellian distribution appropriate to thermodynamic equilibrium, even though the corresponding temperature of the electrons can exceed the temperature of the neutral gas with which the electrons are mixed. This situation is common in laboratory plasmas.

While thermal contact between electrons and heavy particles is poor, collisions do extract energy from the electron gas. Therefore, in the absence of a heat source for the electrons, the electron temperature must ultimately fall to the temperature of the neutral gas. The electrons in the ionosphere are heated as a result of photoionization of the ambient gas by solar ultraviolet radiation. Photoionization constitutes a source of photoelectrons with energies of some tens of electron volts, well in excess of the thermal energies of the other ionospheric constituents. The photoelectrons lose their excess energy quite rapidly as a result of inelastic collisions with the neutral molecules and elastic collisions with the ambient ionospheric electrons. The elastic collisions transfer kinetic energy to the ambient electron gas, thereby heating it. The theoretical calculation of electron temperatures in the ionosphere involves consideration of equilibrium between the rate at which ambient electrons are heated by collisions with photoelectrons and the rate at which they are cooled by collisions with ions and neutral molecules. Preliminary theoretical work on this problem had been done by Drukarev (*22*) as early as 1946, but his contribution was overlooked until quite recently.

The rocket measurements in 1960 were in approximate agreement with theoretical predictions for the region around 200 kilometers but failed to show the predicted decline of electron temperature at high altitudes. However, two of these flights were made in the auroral zone at Fort Churchill and the third, at Wallops Island, was made shortly after a magnetic storm. It was therefore believed that these results were not typical of the normal ionosphere. This belief was reinforced in 1961 when a flight at Wallops Island made during magnetically quiet conditions showed electron temperature declining at high altitudes to a value which appeared to be close to the neutral-gas temperature (*23*).

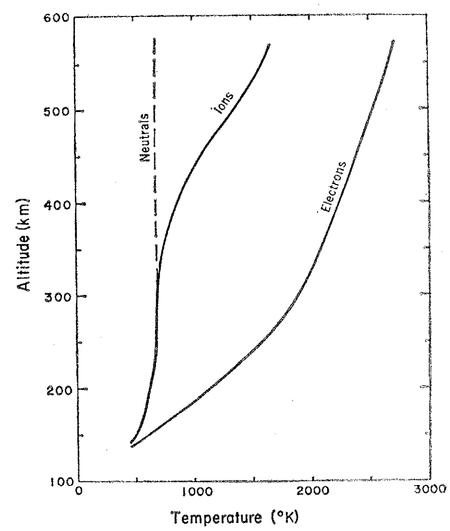


Fig. 3. Average daytime electron and ion temperatures in November 1964, measured at Millstone Hill Observatory by the Thomson scatter technique (*34*). The neutral-particle temperature is inferred.

Radar Measurements

At about the same time, electron temperatures were measured on the Explorer 8 satellite at altitudes above 400 kilometers, and these temperatures also appeared to be close to the neutral-gas temperature (*24*). However, a new experimental technique in which very sensitive radar was used was beginning to provide data on charged particle temperatures (*25*). In the Thomson scatter technique a radar pulse is detected at the ground after reflection by random thermal density fluctuations in the ionosphere. The strength of the returned pulse is proportional to the electron density at the altitude of reflection, and the electron and ion temperatures at this altitude can be determined from the frequency spectrum of the reflected pulse. These measurements can be made only by large and powerful radars at a few specific sites, but the information they provide on ionospheric properties is comprehensive and continuous (*26*).

From measurements made at the Millstone Hill Observatory near Boston, J. V. Evans reported in 1964 (*27*) that the electron temperature exceeded the gas temperature at all altitudes above 200 kilometers and, moreover, that the temperature of the ions was also higher than the temperature of the neutral gas at altitudes above 500 kilometers. This observation was in accord with a theoretical prediction made by Hanson a year earlier (*21*). At high altitudes the ionospheric ions collide much more frequently with ionospheric electrons than with the neutral molecules. As a result, the ions exchange more energy with the electrons than with the neutral molecules and the temperature of the ions approaches the electron temperature. At high altitudes, collisions with the ions constitute the most important means by which the electron gas loses energy. Consequently, as the ion temperature rises the electron temperature rises also (see Fig. 3). It was neglect of this effect, as well as neglect of thermal conduction in the electron gas, which had led to the early prediction that the electron temperature would decrease to the neutral-gas temperature at high altitudes.

Evans's results were soon confirmed by electron temperature measurements made on the Ariel 1 and Explorer 17 satellites, by further rocket flights in the Goddard Space Flight Center–University of Michigan series, and by improved Thomson scatter measurements (*28*). Why, then, had it seemed for a while that the measured electron temperatures at high altitudes were equal to the neutral-gas temperature?

We know, now, that it was our knowledge of the neutral-particle temperatures that was at fault, not our knowledge of the electron temperatures. In the early 1960's available models of the neutral-particle temperature distribution were based largely on satellite drag data obtained for the first satellites. These data reflected a period of particularly high solar activity, when temperatures in the upper atmosphere were between 1500° and 2000°K during the day (see Fig. 2). By 1961 the neutral-particle temperature had declined to about 1000°K, but this change was not yet reflected in the atmospheric models. Misinterpretation resulted from the comparison of 1961 electron temperatures and 1959 neutral-particle temperatures. This tale, incidentally, illustrates an important aspect of upper-atmosphere physics. Because of the great variability of atmospheric properties, comparisons of data acquired at different times are frequently misleading. Simultaneous measurements are needed of related parameters such as electron, ion, and neutral-particle temperatures, but this experimental goal has been achieved only rarely.

An attempt to overcome this difficulty has been made in a theoretical study of the daily variation of the electron temperature in July 1963 (*29*, *30*). The calculations were based on data on the neutral atmosphere, the solar ultraviolet flux, and the ionospheric electron densities, all obtained during that month. In Fig. 4 the theoretical temperatures are compared with temperatures measured during the same month by the Thomson scatter radar at Millstone Hill. The agreement between theory and experiment is good at high altitudes, but there is a significant divergence at the lower altitudes. The reason for this disagreement is not understood.

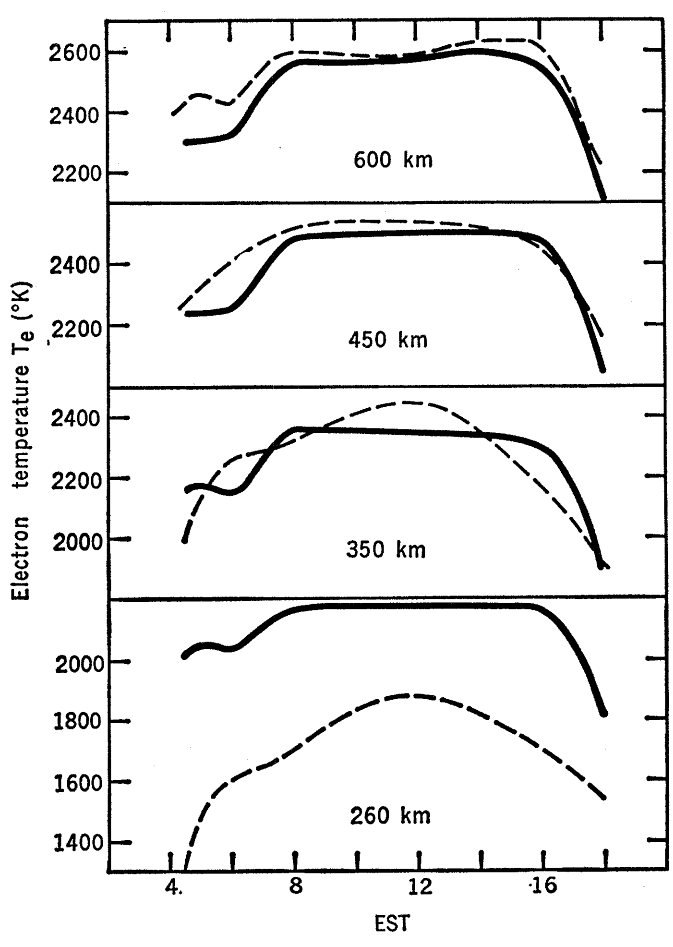
Another discrepancy has been noted at altitudes below 150 kilometers. Thomson scatter experiments indicate, in agreement with theory, that the electron temperature at this altitude is close to the ion and neutral-particle temperatures, but most measurements made with rocket-borne electrostatic probes show electron temperatures to be substantially higher (*31*). It has been suggested that the vibrational temperature of the nitrogen molecules may be as high as 3000°K, and it is possible that energy transfer from nitrogen molecules to ionospheric electrons may be affecting the results (*32*). However, the consequences of such a process have not been worked out in any detail.

Because of the strong interaction between charged particles and magnetic fields, the densities and temperatures of the ions and electrons in the upper atmosphere are influenced by the earth's magnetic field. This geomagnetic control leads to a variation, with latitude, in ionospheric properties which is much more marked than the corresponding variation in the properties of the neutral atmosphere (see Fig. 5). In addition to this variation with latitude, the ionospheric properties vary widely with time of day and with solar activity. In this, the ionosphere resembles the neutral upper atmosphere, but the information on ionospheric variability is, in many respects, less complete than the information on neutral-atmosphere variability.

Conclusion

The upper atmosphere consists of a mixture of gases—electrons, ions, and neutral particles—each of which has a distinct temperature. At most altitudes the electron temperature exceeds the ion temperature and the ion temperature, in turn, exceeds the neutral-particle temperature. Since energy is exchanged between the gases, the different temperatures are related, but the relationship is complex. It is quite common, for example, for the electron temperature to increase while the neutral-particle temperature is decreasing.

We can expect a similar multiplicity of temperatures in the upper atmospheres of the planets. In detail, however, the differences between the atmospheres of the planets must be more striking than the similarities, on account of differences in atmospheric composition and in distance from the sun. It is likely that the absence of a permanent magnetic field on either Mars or Venus causes further substantial differences between the upper atmospheres of these planets and the upper atmosphere of the earth.



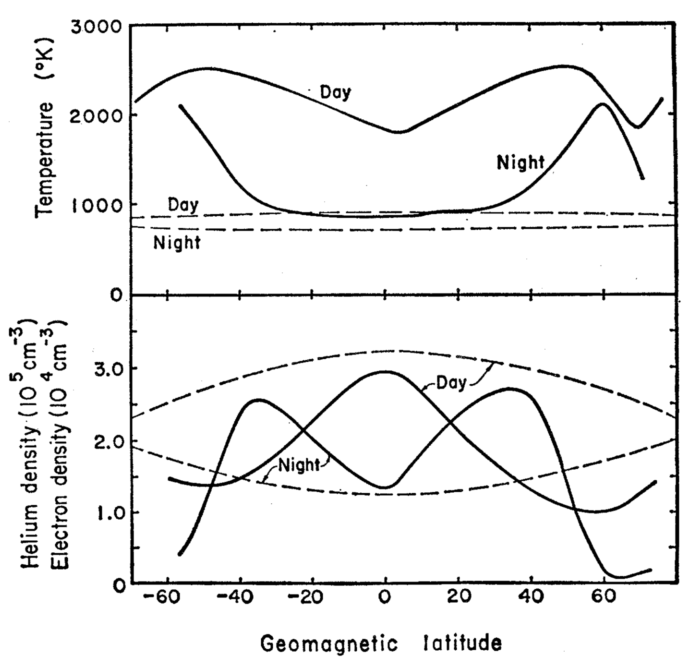


Fig. 4 (left). Dependence on Eastern Standard Time (EST) of electron temperatures at selected altitudes over Millstone Hill in July 1963. (Solid lines) Theoretical values; (broken lines) measured values. [From Dalgarno, McElroy, Rees, and Walker (*30*)]

Fig. 5 (above). Latitude dependence of electron temperature and density at an altitude of 1000 kilometers measured on the Explorer 22 satellite (*35*) during the vernal equinox of 1965 (solid lines). (Dashed lines) Latitude dependence of neutral-particle temperature and helium density, given for purposes of comparison (see *12*).

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Psychochemical Research Studies in Man

Research approaches to the chemistry of the mind of man, although promising, are difficult to interpret.

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Unlike many previous sociologists and historians of science, recent examiners of the scientific process have been struck by the influence of factors within the game of science that appeared to significantly influence trends in research strategies in ways which seemed prepotent over substantive "discoveries" in an area. Kuhn, in his classic essay "The Structure of Scientific Revolutions" (*1*), using the history of some areas of physics as his models, developed a theory of the progression of what he calls "normal science" as the serial emergence and decline of paradigms and, along with them, their practitioners. A paradigm is defined as a shared and consensually agreed upon system of assumptions, acceptable operations, standards for evidence, and rules of conduct for a scientific endeavor that are dominant at a particular time in a field of investigation and is expressed in the form of model problems and solutions. Kuhn would probably regard the subject of this paper as "preparadigmatic"; this status would be earned by both the insufficiency and inchoateness of our information. He pointed out, however, that even in primitive areas such as psychochemistry, rather sizable groups of scientists may often reach temporary agreement about what constitutes good research methodology and acceptable results. Kuhn would predict, however, that consensus in a preparadigmatic field would eventually disappear, and another preparadigmatic school would emerge. The shift, by the biologically oriented student of behavior in man, from electricity to juices in the past decade seems to be a good example of such a preparadigmatic transition. Drawing much of our scientific aura from the basic neurophysiologists, who studied electrical potentials from cells and brain systems in animals in the 1940's and 1950's, we grabbed at any evidence of direct or transduced electricity we could get from the entire surface of man. Taking our cue from such representations of what was *au courant* as Fulton's 1955 *Textbook of Physiology* (*2*) which allowed only 13 pages of talk about neurohumors in 502 pages of brain circuitry, we focused on electricity. The massive number of intervening variables between thought and scalp or palm was

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