SOVIET PHYSICS USPEKHI VOLUME 10, NUMBER 3 NOVEMBER-DECEMBER 1967

525.7

TEMPERATURE OF NEUTRAL AND CHARGED PARTICLES IN THE IONOSPHERE AND THE MAGNETOSPHERE (RESULTS OF MEASUREMENTS WITH ROCKETS AND SATELLITES)*

K. R. GRINGAUZ

Radio Engineering Institute, Academy of Sciences U.S.S.R.

Usp. Fiz. Nauk 92, 207–228 (June, 1967)

I. INTRODUCTION

THE present review of information on the temperatures of neutral particles T_g, ions T_i, and electrons T_e in the earth's ionosphere includes only data obtained with the aid of devices mounted on rockets and satellites, since information concerning earth-based observations (including determination of T_g by measuring the deceleration of artificial satellites) are contained in the simultaneously published review of J. V. Evans [1].

The problem of preparing this review was greatly facilitated by the fact that many good reviews were published recently, devoted to the temperature measurements in the ionosphere: the reviews of Bourdeau [2], Evans [3], Breus and Gdalevich [4], and the temperature section of the report by Champion [5]. Nonetheless, new experimental data, worthy of considerable attention and comparison, were published in 1965–1966.

The temperatures of neutral and charged particles are among the most important characteristics of the ionosphere. Their ratios react sensitively to variations of the remaining parameters of the ionosphere; by comparing the distributions of T_g, T_i, and T_e it is possible to estimate and localize the known sources of heat, and to establish the existence of other sources of heat still unobserved by modern instruments. Therefore the increasing attention paid to temperature investigations of the ionosphere is perfectly natural.

Although the topic and the space allowed for the present review do not make it possible to consider questions involved in the methodology of determining T_g, T_i, and T_e, it is useful to make a few remarks on the role and certain peculiarities of measurements made with the aid of instruments mounted on rockets or satellites.

Land-based measurements of the temperature of the ionosphere (as well as other non-rocket researches) have tremendous advantages over rocket investigations, namely that the data are statistically abundant and relatively cheap to obtain.

It is just the land-based observations of the deceleration of satellites which served as the basis for modern models of the upper atmosphere, including the temperature models. The developing research of the ionosphere by the method of incoherent radio wave scattering, the temperature aspects of which are considered in Evans' review [1], make it possible to determine a large number of ionospheric parameters with increasing assurance of the reliability of their determination. Nonetheless, such measurements have certain shortcomings which are difficult to eliminate. Thus, unless additional data are employed, observations of the deceleration of satellites cannot separate effects produced by the change in the mass from those produced by the change in temperature; it is impossible to record some short-duration state of the medium, for example, it is impossible to determine the temperature of a sufficiently small section of the atmosphere. In other words, the method contains inherently a certain ambiguity and a low resolution, both in time and in altitude. Such features are inherent to some degree also in the method of investigating the ionosphere with the aid of incoherent scattering (see [3]).

By using rocket-borne instruments it is possible in principle to determine the parameters of the medium uniquely and with high resolution in both time and altitude. Therefore, although the number of rocket launchings into the ionosphere are few, owing to the high cost, they are very important, since a comparison of the results of rocket measurements with the results of land-based measurements of the same quantities make the latter unique and provide assurance of the correctness of land-based measurements. In some cases, on the other hand, rocket methods yield information which cannot be obtained at all by other means.

Temperature measurements on rockets and satellites, connected with registration and analysis of the particles having the lowest energies, are among the most difficult rocket experiments, and therefore most experimenters have devoted considerable efforts to ensure and check the reliability of the measured results. Thus, several types of probes were placed on Japanese rockets to measure simultaneously the same quantity [6]; in many rockets launched in the USA, the

*Review article delivered at the Symposium on Physics and Solar-terrestrial Relations, Belgrade, August, 1966.

measurements were made with probes that were separable from the rockets (to reduce the perturbations introduced into the medium by the rocket) [7,8]; identical probes, but differently disposed on the satellite, were used simultaneously (this was done both on the Soviet satellite Kosmos-2 [2] and on the British-American Ariel-1 [10]); rockets traveling relatively close to a satellite were launched in the USA [3]; the data obtained with the aid of rocket probes were compared with data of land-based measurements [12]. The results of the experiments lead to the conclusion that in most cases the information obtained with instruments mounted on rockets and satellites concerning the particle temperatures is sufficiently reliable.

The main experiments on rockets and satellites, which pertain to the distribution of the particle temperature in the ionosphere, were performed after 1960. The main theoretical and semi-empirical models were produced also in the sixties.

According to the Harris-Priester model [11], in which they introduce, besides the ultraviolet radiation from the sun, also a heat source (presumably connected with the solar corpuscular radiation), T_g increases monotonically to the height of the "thermopause," which is the base of the isothermal zone; the height of the thermopause and the value of T_g in the isothermal zone change as functions of the index S (equal to the flux of decimeter radio emission from the sun in $W\text{-}m^{-2}\ sec^{-1}$) and of the local time.

Work on bringing closer together the models of the upper atmosphere and the data observed in experiments are continuing. Figure 1, taken from one of the latest papers [12], shows the dependence of the isothermal zone on S according to Harris-Priester, Yakiya, and Izakov.

The values of T_e in the daytime ionosphere were calculated by Hanson and Johnson [13], Hanson [14], Dalgarno et al. [15], and recently by Geisler and Bowhill [16]. The calculations of [14,15] were published in 1963, when there were still no detailed experimental profiles of $T_e(H)$. The scheme of these calculations is as follows: a) The rates of photoionization and of the photoelectron energy loss processes

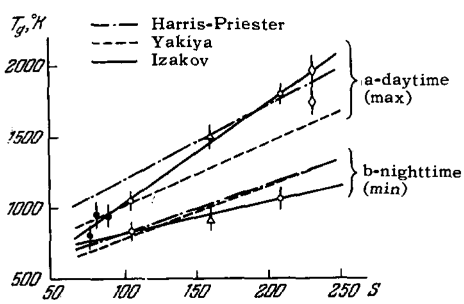

FIG. 1. Dependence of the temperature of the neutral particles T_g in the isothermal region on the flux of solar radio emission S [12].

are examined; these are used to determine the rate of heat influx per unit volume of the electron gas in the ionosphere. b) The rate of cooling of the electron gas (by inelastic collisions with the neutral particles at $H \lesssim 250$ km and by elastic (Coulomb) collisions with ions at $H \gtrsim 250$ km) are considered; Hanson [13] takes into account the thermal conductivity of the electron gas at altitudes $H > 600$ km; Dalgarno et al. neglect the thermal conductivity. In both models [14,15], the rates of heating and cooling of the electron gas are assumed equal at all altitudes, making it possible to determine the distribution of T_e over the altitude. According to both models, T_e begins to increase compared with T_g starting with an altitude of H $H \sim 120$ km, reaching a maximum at $H \sim 220$ km; further increase of H leads to resumption of the thermal equilibrium at $H \sim 350$ km. According to [14,15], the following relation holds when $H \gtrsim 300$ km:

$$T_e - T_g \sim \frac{Q}{n_e^2} T_e^{3/2}.$$

where Q is the influx of heat per cm^3 of electron gas (in $eV\text{-}cm^{-3}\ sec^{-1}$) and n_e is the electron density in cm^{-3}. As regards T_i, according to [14] $T_i \to T_e$ when $H \gtrsim 1000$ km. Bourdeau [2] pointed out the need for theoretically considering the possibility of nonlocal contribution of energy from solar ultraviolet radiation, i.e., the possibility of transfer of absorbed energy at other altitudes. Geisler and Bowhill [16] have shown in their calculations that it is necessary, especially under conditions corresponding to the minimum of solar activity, to take into account the thermal conductivity of the electron gas at altitudes much lower than assumed by Hanson [14]. Allowance for thermal conductivity, i.e., allowance for the possibility of heat transfer at higher altitudes, leads in this case to $T_e(H)$ profiles which differ significantly from those calculated in [13] and [14], in which T_e not only does not decrease with altitude at H = 220 km, but may even have a certain positive gradient. The authors of [16] propose that under the conditions of minimum solar activity, the profiles of $T_e(H)$ should have a form similar to that calculated in [14] and [15]. The question of the agreement between the experimental $T_e(H)$ profiles with the theoretical ones at moderate and low solar activities will be discussed in Sec. III of this review; as to the altitude distributions of $T_e(H)$ corresponding to maximum solar activity, to compare them with the theoretical models it becomes necessary to expect the next maximum of the solar cycle, since, as was already noted, there exist no such experimental data for 1957–1958.

Since 1959, when experiments with charged-particle traps mounted on Soviet lunar rockets have reliably established the existence of a plasma sheath around the earth, with thermal velocities of the ions at altitudes reaching ~20 000 km (Gringauz, Besrukikh, et al. [17]), this region has not been given a universally accepted name. Thus, for example, in different articles it is called "the ionized component of the geocorona" (Gringauz et al.[18]), the "protonosphere" (Geisler and Bowhill [16,19]), the "magneto-ionosphere" (Taylor et al.[20]), "plasmasphere" (Carpenter [21]), etc. Apparently the most suitable name for this zone is "peripheral region of the ionosphere"; this is will be used in the present review.

According to the theoretical estimates of Geisler and Bowhill [19], at heights considerably larger than 1000 km, we have $T_i = T_e = T$ (this is also obtained in [14]), but the value of T is larger than in [14]; at a distance of 8000 km along the magnetic force line crossing the $H = 1000$ km level at the geomagnetic latitude $L = 40°$, depending on the phase of the solar-activity cycle (according to [19]), T should range from 3000 to 3400°.

In concluding this section it should be noted that all the theoretical calculations of the charged-particle temperatures in the ionosphere, reported in [13 16], include stages in which certain not fully reliable estimates are used (for example, the determination of the rates of photoelectron generation and of their mean energies at different altitudes). In addition, there is practically no doubt that there exist additional sources of heating of charged particles (fluxes of nonthermal particles, hydromagnetic waves, electric fields), which are not taken into account in the theoretical calculations of [14,17]. There is therefore no reasons for always expecting the experimental results to agree with the calculations; a theoretical model of the charged-particle temperature distribution can be regarded as fully satisfactory if it qualitatively agrees with experimental data and gives approximately close quantitative results.

II. MEASUREMENTS OF T_g WITH THE AID OF DEVICES MOUNTED ON ROCKETS AND SATELLITES

The neutral-particle temperature T_g in the upper atmosphere is determine essentially by one of two methods. The first consists in the following: The altitude distribution of the density is first determined, and T_g is then calculated from the height of the homogeneous atmosphere (altitude scale). The other method consists in expelling a chemical reagent from the rocket into the investigated atmosphere (Na, K, AlO) and measuring the Doppler broadening of the resonant radiation of its particles, produced under the influence of the solar radiation.

Both methods give data that are averaged over the altitude; in the second case, the averaging interval is determined by the dimensions of the luminous cloud (since the luminosity observed on earth is determined not only by the particles of the surface layer of the cloud, but also by the particles in its entire thickness along the line of sight to the observer).

The altitude scale ($\overline{H} = RT_g/Mg$) depends essentially on the average molecular weight M; therefore for an exact determination of T_g it is necessary to know the mass spectrum of the particles in the investigated region. If H is determined for one component of the mixture of neutral gases from mass-spectroscopic measurements, then the value of T_g is determined with higher accuracy than from data obtained by altitude measurements of the atmosphere density with the aid of manometers mounted on rockets (let alone the data based on analysis of satellite deceleration).

The continuous curves in Fig. 2 show the values of T_g determined from data of Soviet mass-spectrometric measurements. These results were obtained by launching geophysical rockets at medium latitude of the USSR (Pokhunkov [22,23]). The value of T_g was determined by two methods: (a) by measuring the altitude variation of the relative concentrations of two inert gases, and (b) by determining the altitude distribution of the partial pressure of one component of the mixture of neutral gases. To measure the temperature by method (a) it is necessary that a stable gravitational separation exist in the atmosphere; according to rocket experiments with mass spectrometers, this exists at altitudes $\gtrsim$ 110 km. The errors in the determination of T_g are estimated by the author of the experiments to be 10% of the measured values. In all three experiments, it is seen that T_g increases starting with $H \sim 100$ km. The first two measurements correspond to an average level of solar activity (the flux of radio emission from the sun at wavelength $\lambda = 10.7$ cm is $S = 175$ in units of 10^{-22} W/m^{2}Hz); in the third measurement (1961), $S = 100$. According to these data, the thermopause (lower limit of the isothermal zone) is located much higher than 200 km (during the time of the experiment of 15 November 1961 its altitude was > 300 km), and T_g in the isothermal zone is $\sim 1500°$. We note that the existence of the isothermal zone is seen only on the curve of 15 November 1961; the first two measurements were apparently made at insufficient altitudes.

Blamont, Lory, and their co-workers, starting with 1960 [24,25,26], performed a number of experiments aimed at determining T_g. Chemical reagents were expelled from rockets launched in Algiers and in the USA during the sunrise and sunset twilight, and a study was made of the broadening of the resonance lines of K and Na and of the AlO bands, which fluoresce under the influence of solar radiation. The same Fig. 2 shows some results obtained for T_g by Blamont [25]. The number of measurements performed by Blamont's group at altitudes > 200 km is also small and insufficient to confirm the existence of an isothermal zone and determine its temperature and height of the thermal pause. At altitudes from ~210 to ~410 km, there are no measurements on the 1961

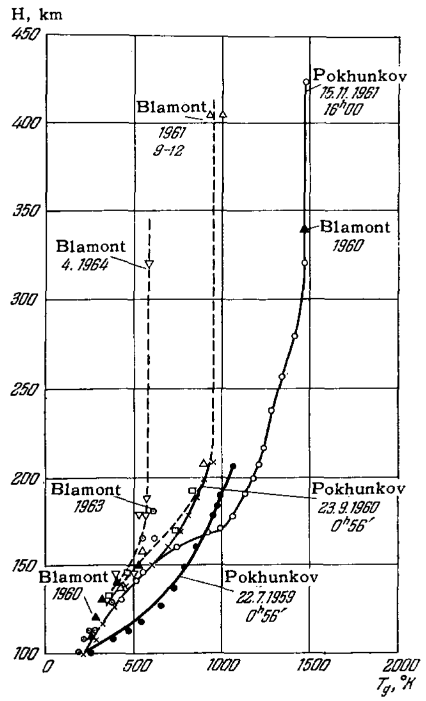

FIG. 2. Temperature of neutral particles as measured by Pokhunkov and by Blamont et al.[22-25]

curve, and if this zone contained a minimum or maximum of T_g, the measurements did not reveal it. All the data pertaining to altitudes $\lesssim$ 150 km, obtained in different years by the mass-spectrometric method and by measuring the broadening of the resonance luminescence lines of K and Na, are relatively close to each other and give the impression of a small dependence of T_g at these altitudes on the phase of the solar cycle. Although it follows from the plot of the data of Blamont et al., dating back to 1961, that the thermopause is located in these measurements at a lower altitude than follows from the data of Pokhunkov, which also were obtained in the fall of 1961, and the temperature above it is lower, it must be borne in mind that these measurements were carried out in different geographic regions and, what is particularly important, at essentially different local times (the measurements on the Soviet rocket of 1961 were carried out at 16^h local time.

Spencer, Brace, et al.[27] determined T_g by a device that became detached from four rockets launched in 1962-1964 over Wallops Island (Virginia, USA) during different times of the day, daytime, during sunset, at night, and also during the solar eclipse of 20 June 1963. These devices were called by them "thermospheric probes" and contained an omegatron mass spectrometer, tuned to definite partial concentrations of one component of the neutral atmosphere (N_2), as well as Langmuir probes. These exceedingly interesting experiments have revealed good agreement between the experimentally determined $T_g = T_{N_2}$ with the values of T_g determined by the Harris-Priester model [11]. Figure 3 shows $T_g(H)$ profiles obtained during the time of these four experiments (data on the measurements of T_e, obtained during the time of the same experiments, are given in Sec. IV of the present review). In all four $T_g(H)$ profiles one can see clearly the existence of an isothermal zone; the height of the thermal pause for the local time $\sim 16^h$ is smaller than in the results shown in Fig. 1 pertaining to 1961 - this is natural, since it reflects the smaller influx of heat in the upper atmosphere in 1963, compared with 1961; for the same reason, the height of the thermal pause is lower during sunset than in daylight.

Although the data presented on rocket measurements by Pokhunkov, Blamont, Spencer, Brace, et al., are certainly not equivalent from the point of view of the statistics to the results obtained for T_g by observing the deceleration of satellites, they are very important because of the large time resolution, locality, and uniqueness of interpretation of their results.

The rocket-measurement data presented above either confirm the existence of an isothermal zone at altitudes larger than 200—300 km, or do not contradict its existence. However, the published values of T_g include also some that indicate a nonmonotonic variation of T_g at the indicated altitudes. Thus, Mikhnevich[28] concluded, after reducing results of manometric measurements performed on geophysical rockets rising to 450 km, that during the time of the measurements the value of T_g had minima at altitudes from 200 to 300 km.

It seems to us that the data gathered todate on the T_g profiles are insufficient to negate the possibility of at least a short-duration existence of altitude distributions similar to those presented in [28].

T_g was also determined by instruments mounted on the American aeronomical satellite Explorer-17, which was launched in April 1963 on an eccentric orbit (perigee 258 km, apogee 420 km). Newton et al determined T_g from altitude measurements of the density, which was measured on this satellite with the aid of manometers; the average mass of the particles

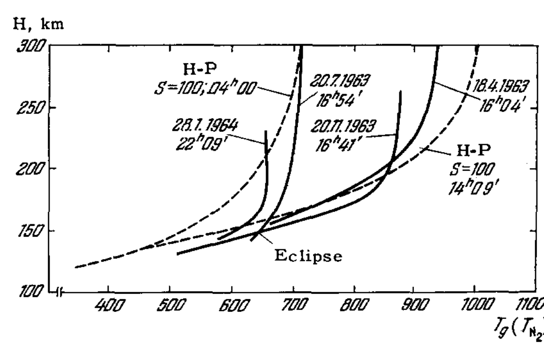

FIG. 3. Altitude distribution of $T_g(H)$ obtained during four launchings of rockets in 1963-1964 by Spencer et al.[27]

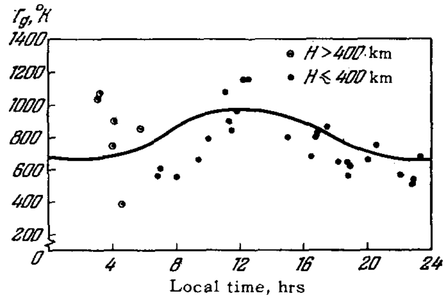

FIG. 4. T_g as measured on Explorer-17 with the aid of manometers [29]. The solid curve shows the values of T_g in accordance with the Harris-Priester model for S = 90.

was determined from mass-spectrometer data [29]. According to the estimate of the authors of the experiment, the error in the determination of T_g is probably lower than ±20%. Figure 4 shows a plot of the values of T_g, obtained at different local times (black dots at altitudes $H \lesssim 400$ km, circles at altitudes $H > 400$ km). The solid curve shows the dependence of T_g on the local time, corresponding to the Harris-Priester model for S = 90. The appreciable scatter of the points in the interval from 3 to 7 hours offers evidence of a large scatter of the altitude scale. The authors of [29] believe that the causes of this scatter are not clear - it may be connected with variations of the temperature or of the average particle mass or both, or else with the violation of the diffusion equilibrium. The values of T_g vary approximately by a factor of 2 (from ~500 to ~1000°). It is noted in [29] that an analysis of the results obtained in the altitude interval 500—600 km continues, in order to separate the effects connected with changes in the particle mass from effects connected with temperature variations.

All the T_g data presented above were obtained by reducing the results of such measurements of the physical characteristics (altitude scale, Doppler broadening of the resonance line) which are connected with averaging over relatively large regions of the ionosphere. Yet it is possible in principle to make direct absolute measurements of T_g which are much more localized.

In 1961 such a possibility was pointed out by Pressman and Yatsenko [30], who proposed to use for this purpose measurements of particle fluxes flowing into two narrow tubes located at different angles to the velocity vector of the space vehicle, and who presented an appropriate theory.* The difficulties in the realization of such measurements are purely technical and are connected with the problems involved in determining accurately the orientation of the space vehicle and measurement of very small currents. Analogous difficulties are encountered in the realization of another closely-related method of measuring T_g, called by Spencer, Brace, et al. the "velocity scanning method" and consisting of determining T_g from the depth of modulation of the current of the omega mass spectrometer, due to the periodic variations of the orientation of its input aperture as a result of rotation of the thermospheric probe [27]. The authors of [27] attempted to use the experimentally observed modulation of the mass-spectrometer current on thermophysical probes to effect an absolute measurement of T_g; the preliminary results published to date are apparently somewhat overvalued. There is no doubt, however, that such methods are quite promising. There is every reason to expect the technical difficulties to be overcome and the reliability, and especially the altitude resolution, with which T_g is determined to become much higher.

*The author of the present review has learned, after the end of the Belgrade symposium, of a paper in which a similar method was proposed earlier [58].

III. MEASUREMENTS OF T_i

The number of reported attempts of measuring T_i directly with instruments mounted on rockets and satellites is small; the number of successful attempts is even smaller.

Sharp, Hanson, and McKibbin performed an experiment aimed at directly determining T_i by measuring the ion velocity distribution using the retarded-potential method and a flat ion trap [31]. Such traps were mounted on two satellites, launched in 1961 and 1962 on nearly-circular orbits, with large inclinations to the equator. The directions of the normals to the trap were close to the directions of the satellite-velocity vectors. On the first satellite (altitudes from ~230 to ~240 km), the calculated values of T_i fluctuated from one cycle of changes in the retarding potential to the other, within a range from ~1200 to ~2400°; in the second (altitude H ~ 245 – 280 km), T_i fluctuated from ~600 to ~ 1800; with this, in some cases the values of T_i turned out to be lower than the expected values of T_g. The authors note that since the instrument should in principle be an excellent means of measuring T_i, the results are highly disappointing.

Three-electrode ion traps of the honeycomb type were installed on the satellite Kosmos-2 (April 1962, Gringauz et al. [9], Afonin et al. [32]) for the measurement of T_i; in these traps, the external grids were replaced by fittings consisting of parallel tubes which were long compared with their transverse dimension. The principle of measuring T_i by such a trap is similar to the principle of measuring T_g with the aid of a narrow tube, as mentioned at the end of the preceding section. Since the Kosmos-2 satellite executed a complicated rotation, the determination of T_i with the aid of the honeycomb traps were pos-

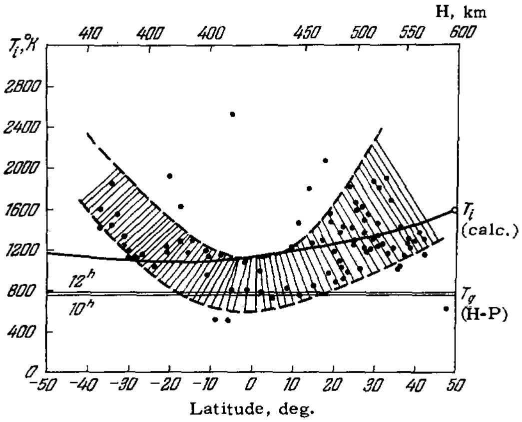

FIG. 5. Results of measurements of T_i over one month with the aid of a charged-particle trap mounted on Ariel-1 [33].

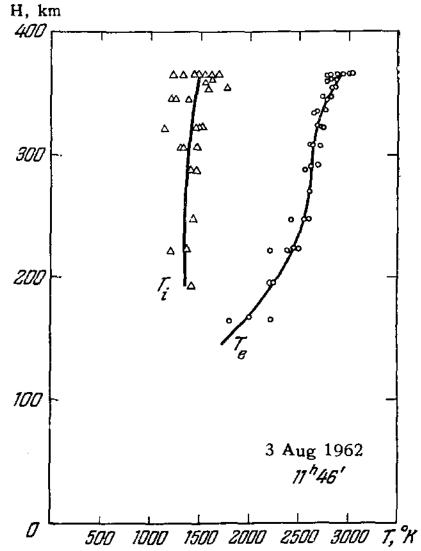

FIG. 6. $T_i(H)$ and $T_e(H)$ profiles corresponding to noontime in August 1962.[34]

sible only in a few cases, when the orientation of the satellite was suitable for this purpose. The measured values of T_i pertain to the daytime and to altitudes < 400 km. They are approximately half as large as the values of T_e measured simultaneously with the aid of Langmuir probes; thus, $T_i = 1300° \pm 200°$ for H = 260 km and $1500° \pm 200°$ for H = 300 km.

The values of T_i were determined from data of a spherical ion trap mounted on the British-American satellite Ariel-1 launched in April 1962 (Boyd and Rait [33]). The values were determined from the width of the peak of the plot of the second derivative of the collector current, corresponding to the O^+ ions in the altitude interval 400–600 km, from data obtained approximately over one month; the local solar time varied within a range of two hours. The measured values of T_i are shown in Fig. 5. According to the authors' estimate, the error of each individual determination does not exceed 200°. Therefore the scatter of the points apparently reflects real differences in the measured temperatures. They are partly due to the fact that the measurements were made at different local solar times, and partly connected with seasonal variations of T_i and with variations of the magnetic activity. The main conclusion is that there exist real variations of T_i of oxygen ions from day to day, amounting to several hundred degrees, and there is a tendency for T_{iO^+} to increase with increasing latitude. The two lines parallel to the abscissa axis in Fig. 5 represent the values of T_g in accordance with the Harris-Priester model for local time 10^h and 12^h, while the continuous curve represents the values of T_i calculated by Wilmor from the T_i data obtained on Ariel-1, under the assumption that the neutral gas is heated by the electrons via Coulomb interactions between the electrons and the ions. The authors of [32] note that since the rate of energy exchange between particles depends noticeably on the mass of the ions, the experimentally observed scatter of values of T_i can be connected with variations of the ionic composition.

The altitude profile $T_i(H)$ was obtained (simultaneously with the $T_e(H)$) profile) by Nagy et al.[34] at altitudes from 180 to ~ 365 km during daytime on 3 August 1962 with the aid of a device, separated from the rocket, consisting of a spherical ion trap and a Langmuir probe (Fig. 6). The entire indicated interval is isothermal from the point of view of T_i ($T_i \sim 1800°$): T_e increased monotonically with altitude, exceeding T_i appreciably at all altitudes. Attention must be called to the fact that the form of $T_e(H)$ in Fig. 6 greatly contradicts the theoretical models of Hanson [13] and Dalgarno [14] (T_e has no maximum near H ~ 220 km).

It seems somewhat strange that no further attempts to measure T_i by direct methods at altitudes H < 1000 km were made since 1962, judging from the published data.

Inasmuch as we know, the first rough estimate of T_i in the peripheral region of the ionosphere (H $\lesssim$ 20 000 km) was made by Gringauz, Bezrukihk, et al. in [17], where it was indicated that T_i does not exceed several times 10 000 degrees.

Gringauz, Bezrukikh, and Breus [35] proposed to use, for the estimate of T_i, modulation of the collector current in an iron trap with zero potential on the outer grid, resulting from rotation of the spacecraft. The application of this method to certain results obtained on the satellite Elektron-2 [36] has made it possible to lower the upper limit of the possible values of T_i at altitudes ~5000–7500 km to 9000–10 000°; actually, T_i is apparently much lower than the indicated value at these altitudes.

IV. MEASUREMENT OF T_e

A. Vertical Distribution of T_e (Rocket Measurements of T_e)

The most reliable determinations of the vertical distribution of T_e are obtained, of course, during measurements carried out when rockets are launched on trajectories that are close to vertical. Among the first published results of rocket measurements of T_e are those obtained in Japan (Aono, Hirao, Miyasaki [37,6]) and in the USA (Brace, Spencer, et al.[7,8]). The aggregate of these results has shown that although the measured values of T_e are sometimes low at altitudes $H \lesssim 150$ km ($\sim$1000—1200°), and are apparently close to T_g, nevertheless at other times one observes at these altitudes also high values of T_e (up to $\sim$2000°K), thus evidencing undisputed absence of thermal equilibrium during the time of these measurements; T_e increased with increasing altitude, reaching values close to 3000° in daytime experiments [7,8]. Subsequently, such high values of T_e were observed many times in experiments performed both on rockets and on satellites.

The absence of thermal equilibrium in the F region of the ionosphere was quite clearly revealed by the experiments of Spencer et al. with the aid of thermospheric probes [8]; these experiments were already mentioned in Sec. II. Figure 7 shows the results of two measurements made in 1963 with two rockets launched from Wollops Island (Virginia, USA). Both rockets were launched at the same local time ($\sim 16^h - 17^h$), the first on 18 April and the second on 20 July. The figure shows the data obtained during the launching of the rockets. We shall consider first the data of 18 April. A comparison of the measured values of T_e and T_g demonstrates convincingly the absence of thermal equilibrium in the F region of the ionosphere; at H = 250 km the value $T_e \sim 2000°$ is more than double the value of T_g. The second pair of diagrams, which show the results of the measurements of 20 July 1963, reflect the state of the ionosphere during the time of an eclipse. During the time of the rocket travel, the covered part of the surface of the solar photosphere was 85 to 75% of its total surface.

These measurements, together with those made on rockets launched during the time of the same eclipse at Ft. Churchill (Smith [38]), present remarkable experimental evidence that the ultraviolet radiation of the sun is the principle source of electron heating in the F region of the ionosphere.

If we compare the values of T_e and T_g in Fig. 7, corresponding to 18 April and 20 July, then we can see that T_e during the eclipse is only $\sim 1/2$ of the value of T_e under normal conditions. We note that T_g during the time of the eclipse differs from the value of T_g of 18 April much less, although it is also somewhat lower. At altitudes below 150 km, the differences between the values of T_e on 18 April and on 20 July are insignificant (in both cases $T_e \sim 1000°$). Both the differences between the T_e curves at sections higher than $\sim$150 km, and their similarity at lower latitudes, are not the results of seasonal variations (April—July). This is seen from the fact that a similar effect was observed by Smith et al. from data of Langmuir probes mounted on four rockets launched within approximately one hour during the time of the same solar eclipse of 20 July 1963 (the maximum eclipse was at $21^h06'$ UT), and after the eclipse at Ft. Churchill (Fig. 8). The extreme right curve corresponds to the rocket launched after the end of the eclipse. Below H = 150 km, the value of T_e changes little at the end of the eclipse (the measurement error is ±100°K), and increases sharply in the altitude interval 150—190 km.

The measurements of [8,38] show quite clearly that, as noted above, the solar ultraviolet radiation is the

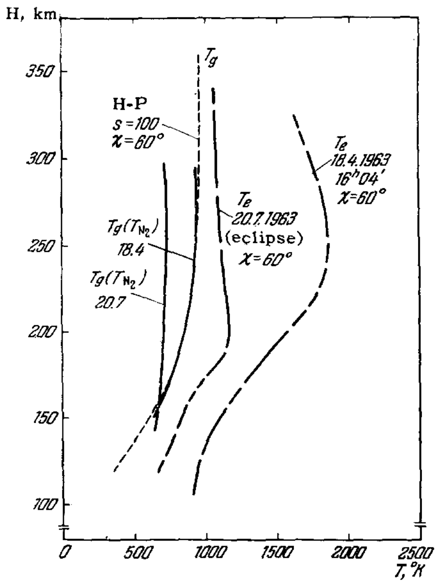

FIG. 7. T_e(H) and T_g(H) obtained during the time of two rocket launchings in 1963.[8]

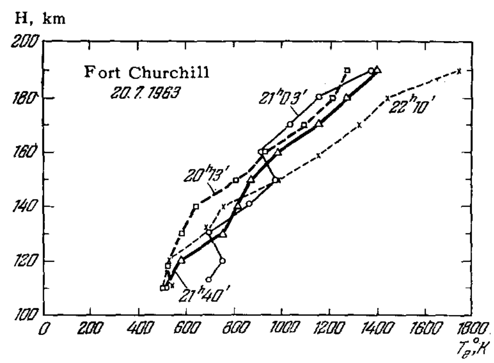

FIG. 8. T_e(H) profiles obtained from four rocket launchings over Fort Churchill during the eclipse of 20 July 1963 by Smith et al.[38]

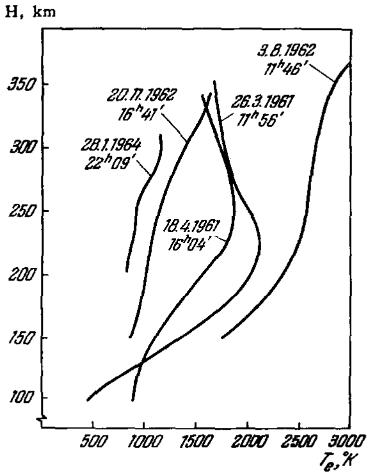

FIG. 9. $T_e(H)$ profiles obtained by a group headed by Brace and Spencer in 1961-1964. The curve for 3 August 1962 was obtained by Nagy [34].

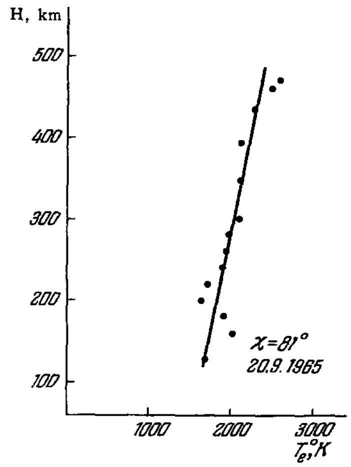

FIG. 11. $T_e(H)$ profile obtained with the aid of Langmuir probes mounted on a rocket by Gdalevich, Gringauz, et al. [41] in morning hours in September 1965 at medium latitudes.

principal source of electron heating in the F region of the ionosphere, and also that this radiation is not the main source of electron heating in the E region, in which, undoubtedly, there is some other source (for example, the electric field produced by the dynamo effect).

At the same time, the dependence of n_e in the E region on the far shortwave radiation of the sun has been well known for a long time, both from the diurnal variation of the critical frequencies f_{OE} as obtained from data of ionospheric stations, and from the ionospheric experiments during the time of solar eclipses (see, for example, N. D. Papaleksi [39], 1938). This dependence came fully into play during the time of the indicated rocket experiments of Smith et al.[38] (who reached the conclusion that the daytime E region is produced essentially by x rays from the sun). Thus, the solar radiation, being the main source of ionization of the E region, is not the main source of electron heating of this region.

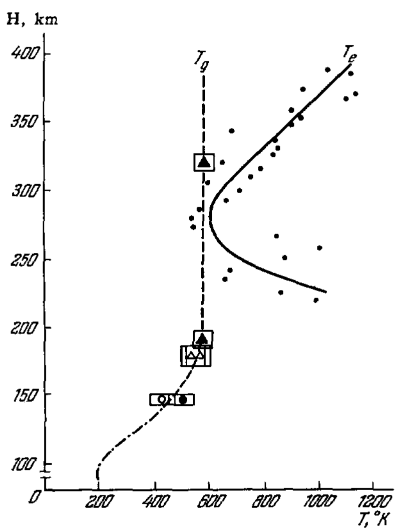

FIG. 10. Results of simultaneous measurements of T_e, carried out by Bourdeau, and of T_g, carried out by Blamont, during the time of two satellite launchings during twilight in the Sahara [26].

Of considerable interest are the results of the rocket experiments aimed at measuring T_e, carried out by Ulwick, Pfister, et al. with rockets launched in Ft. Churchill on 8 February 1964 directly into the region of the visible aurora [40]. At altitudes 300—320 km, values $T_e \sim 5000°$ were observed, much higher than the values of T_e usually observed in the ionosphere.

The great variety of the $T_e(H)$ profiles obtained at medium latitudes by the group of Spencer and Brace in 1961—1964 is illustrated in Fig. 9. The same figure shows the $T_e(H)$ curve (Nagy et al.) shown earlier in Fig. 6. Besides the daytime results, which are similar to the theoretical models of Hanson and Dalgarno [14,15]*, daytime results were obtained contradicting these models, and also $T_e(H)$ profiles pertaining to different times of the day, for which no theoretical calculations have been made as yet.

Interesting results were reported in 1965 by Blamont et al.[26] on the measurement of T_g by means of the resonant luminescence of K and AlO, and on the measurements of T_e performed by Bourdeau with the aid of Langmuir probes during the time of launching of two rockets in Sahara in 1964 in twilight (Fig. 10), which revealed a deep minimum at 275 km.

Langmuir probes on a geophysical rocket launched in the morning hours to an altitude ~500 km in medium latitudes of SSSR in September 1965 yielded the $T_e(H)$ profile shown in Fig. 11 (Gringauz,

*It is reported in the Cospar Information Bulletin (27, 115, 1965) that a maximum of T_e was observed at $H \approx 300$ km with the aid of the L-3-1 rocket launched in Japan on 11 July 1964 to a height of $H = 850$ km at $11^h50'$ local time.

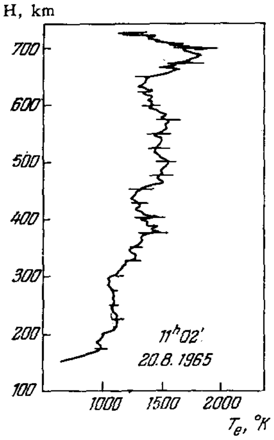

FIG. 12. $T_e(H)$ profile obtained by Hirao with a Japanese rocket. It shows clearly the alternation of maxima and minima of T_e at different altitudes [43].

Gdalevich, et al.[41]). The plot was obtained during the rise of the rocket; each point is the result of averaging of the values of T_e obtained in an altitude interval of ~50 km. The solid curve shows an undisputed tendency for T_e to increase with altitude, although the arrangement of the points has an oscillating character. These results are reported in greater detail in a paper by Gdalevich et al., presented at the present symposium. The results of the Ariel-1 satellite [42], point to an increase of T_e with altitude at heights larger than 400 km in 1962 [42].

As already noted in the introduction, Geisler and Bowhill [16] have shown that a proper allowance for the thermal conductivity of the electron gas makes it possible to obtain a greater variety of $T_e(H)$ profiles than afforded by the earlier models [14,15]. They noted that the absence of a maximum of T_e at altitudes ~220 km may be possibly due to insufficient efficiency of cooling of the electron gas at high altitudes under the conditions of the minimum of solar activity, in connection with the appreciable decrease of the electron concentration during that period, and the corresponding decrease in the efficiency of electron cooling by heat transfer to other particles.

Interesting results of the measurements of T_e, carried out in Japan in August 1965 with rockets launched at approximately 11^h local time to a height of more than 700 km, were reported by Hirao [43]. The measurements, performed by the high frequency probe methods, have shown that besides having a general tendency to increase with altitude, the $T_e(H)$ distribution has a number of minima and maxima, and that T_e has a unique "layered" structure with "layer" thickness (defined as the distance between the neighboring maxima or minima) on the order of 100–150 km (Fig. 12). A similar form of $T_e(H)$ was obtained also by another Japanese rocket launched in 1965 to an altitude more than 300 km [43]. The author advances considerations in favor of the premise that such altitude oscillations of T_e are due not to oscillations of the intensity of the heat source Q, but oscillations in the heat loss, which are apparently connected with altitude variations of the concentration and chemical composition of the ions and neutral particles. No arguments of aeronomic character favoring the existence of such variations are presented by the author. One can note, however, that the oscillatory character of the arrangement of the points on the preceding figure (see Fig. 11) recalls somewhat the results reported by Hirao. Thus, part of the results of the rocket measurements of T_e do not agree with the theoretical models of $T_e(H)$ produced to date, and further experimental and theoretical investigations of this problem are necessary.

Although rocket measurements of T_e in the ionosphere were made in several regions of the world (Japan [6,43], USA [7,8,34], Canada [38,40], Algiers [26], USSR [41]), the fact that the measurements were not made simultaneously and that their number is not large makes it impossible, in our opinion, to draw from them any conclusions concerning the dependence of T_e on the latitude, on the local time, etc. Such conclusions can be drawn on the basis of satellite measurements.

B. Dependence of T_e on the Local Time, Latitude, etc. (Measurements of T_e on Satellites)

The table lists data on certain satellites used to measure T_e by probe methods.

Probe measurements of T_e with the Explorer-8 satellites were made by Bourdeau et al.[44,45]. The eccentricity of the orbit and the fact that the measurements were made only during the time when direct

Name	Launching date	Perigee, km	Apogee, km	Inclination to equator
Explorer-8	3.11 1960	425	2 400	50°
Kosmos-2	7.04 1962	212	1 540	49
Ariel-1	26.04 1962	400	1 200	54
Ariel-1	June 1962	260	317	75
Ariel-1	July 1962	160	181	75
Explorer-17	3.04 1963	258	920	58
Explorer-22	9.10 1964	1000	1 000	80
IMP-2	4.10 1964	200	95 016	33,53°

radio communication with the earth was possible, has made it necessary for the authors of the experiment to analyze the results by making the simplifying assumption that T_e is independent of the altitude. The most important and reliable results of the experiment was the observation of a considerable peak of T_e near the sunrise time (up to 2.5 T_g).

The satellite Kosmos-2 was used to measure T_e in the altitude interval from 212 to 550 km only during the daytime and when direct radio communication between the satellite and the earth was possible (Gringauz, Gorozhankiĭ, et al.[9], Alfonin et al.[32]). The insufficient amount of data on T_e did not give the authors of the experiments any grounds for attempting to separate the influence exerted on T_e by latitudinal and by other variations. The measured daytime values of T_e in the region F of the ionosphere ranged from 1800 to ~3000°; at those points where the values of T_i were measured simultaneously, the values of T_e exceeded T_i by 2—2.5 times, thus indicating absence of thermal equilibrium.

The measurements of T_e on the satellite Ariel-1 were made with the aid of two flat Langmuir probes; the results of the measurements were memorized along the entire satellite orbit (Bowen, Boyd, et al.[42], Willmore[46]). This yielded a large amount of data, but simultaneous measurements of the altitude, latitude, and local solar time along the orbit of the satellite made the separation of the influence exerted on T_e of each of the indicated factors a difficult task. The authors of [42] and [46], to study the influence of these factors separately, subjected the primary results to a complicated statistical analysis, assuming the seasonal variations of T_e during the four-months' period to be negligibly small. Data of T_e pertaining to individual revolutions of the satellite, were not published. From the published results of the reduction of the primary measurement data on T_e, obtained on Ariel-1 during the period from 28 April to 22 August 1962,[42,46] it can be seen that T_e increases with altitude in the entire interval of the investigated altitudes and at any local time (this does not agree with the theoretical models of Hanson and Dalgarno[14,15], according to which there should be no increase of T_e with altitude in the daytime ionosphere at these altitudes, but can be reconciled with the model of Geisler and Bowhill[16]), and that T_e also increases with increasing geomagnetic latitude. It is noted in [42] that an increase in the intensity of the ultraviolet solar radiation (an index of which is the increase of S) causes an increase of T_e. In all cases it is noted that a negative correlation exists between T_e and the electron density n_e (and is fully explicable from the point of view of the Hanson and Dalgardo theory, in connection with the changes in the conditions of cooling of the electrons with decreasing n_e).

It is somewhat strange that the largest registered values of T_e reported in the published results of the measurement of T_e with Ariel-1 are values slightly exceeding 2000°K, whereas rocket measurements (see the preceding section) and measurements with other satellites frequently gave $T_e \sim 3000°$. We shall return to this question later.

The diurnal variation of T_e at different altitudes did not contain, according to [42], the "peak" observed with the Explorer-8 near sunrise, but subsequently Willmore[46] again reviewed these results and found the indicated peak of T_e. According to [46], the magnitude of this peak decreases with altitude (Fig. 13). The authors of [42] and [46] observed an appreciable dependence of the distribution of T_e at altitudes 400—1200 km on the geomagnetic field; the increase of n_e during the time of magnetic storms is always accompanied by a decrease of T_e; changes of T_e in magnetic storms occur along the magnetic shells. (We note that according to earlier rocket measurements of Brace and Spencer[7,8] the magnetic disturbances cause not a decrease but an increase of T_e.) Near the geomagnetic latitude 50° there is noted a weakly pronounced maximum of T_e.

Willmore[46] notes also that the negative correlation of T_e and n_e are always observed during nighttime, and also the change in the altitude gradient of T_e at ~600 km, near which the ionic composition changes, show that these changes are connected with changes in the rate of cooling of the electrons by collision, and thus point to the existence of an active heating mechanism under nighttime conditions. These indications are particularly strongly pronounced at lattitudes larger than 30°.

During the period close to the time when the measurements of T_e were made with Ariel-1, comparison of the short-duration measurements were made at smaller altitudes with two satellites, the approximate orbit parameters of which are indicated in the table (Sagalyn, Smiddy, and Bhargava[47]). The authors of [47] do not present exact measurement dates, indicating only that they pertain to the time interval June - July 1962. These measurements, carried out with the aid of spherical probes screened by grids, also revealed a clear-cut diurnal variation of T_e, with a peak near sunrise. The daytime values of T_e at altitudes 250—300 km were ~3000°.

Brace, Spencer, and Dalgarno published in 1965 part of measurement results obtained with the aid of cylindrical Langmuir probes on the Explorer-17 satellite[48]. This part of the results pertains essentially to the period from 4 April to 10 July 1963, to geographic latitudes from −30 to −50°, and to the altitude interval from −260 to ~550 km. In addition, certain data are presented corresponding to the geomagnetic latitudes 10 and 60°.

In connection with the excentricity of the orbit of Explorer-17, the authors of [48] resort, in the analysis aimed at revealing the dependence of T_e on various

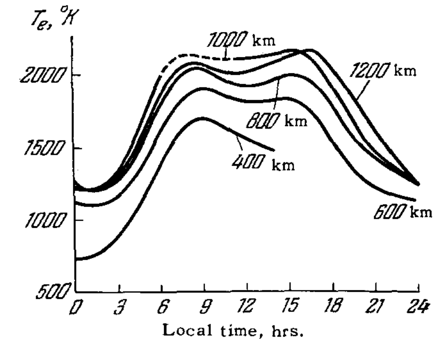

FIG. 13. T_e measured at different altitudes and at different local times with the aid of the satellite Ariel-1 [46].

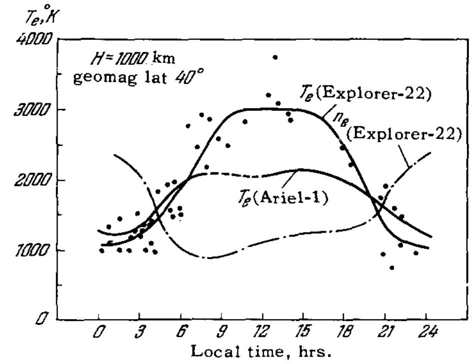

FIG. 15. Diurnal variations of T_e observed on Explorer-22 [49] and Ariel-1 [45], and also of n_e measured on Explorer-22 [49]. All data pertain to H = 1000 km and 40° geomagnetic latitude.

factors, to simplifying assumptions. The main conclusions concerning the distribution of T_e over the altitude and the geomagnetic latitudes, concerning the negative correlation of T_e and n_e, concerning the existence of a nighttime source of heating the ionosphere, agree qualitatively with the deductions of the authors of the measurements with Ariel-1. There are, however, certain quantitative discrepancies. Thus, the morning peak of T_e is ~2700°, which is much higher than the value determined on Ariel. Noting, as Willmore did, the need for the existence of an energy source producing the nighttime difference between T_e and T_g, the authors of [48] assume that to explain the larger values of T_e measured with Explorer-17 at ~400 km it is necessary to have an influx of heat ~20 eV-cm^{-3}sec^{-1}, which is five times the amount required to explain the data of Ariel-1, and that the flux of electrons with energy ~100 eV, corresponding to the energy flux 1×10^{-2} erg-cm^{-2}sec^{-1}, at a heating efficiency 0.1, could explain the observed value of T_e without contradicting the data obtained in other geophysical observations.

The Explorer-22 satellite, launched in 1964, had a circular orbit with large inclination (~80°). The almost complete absence of changes in the altitude and the rapid variations of the latitude at slow variations of the longitude make it practically an ideal apparatus for the study of latitude variations of the ionospheric parameters. The first results of the measurements, obtained on this satellite with the aid of Langmuir probes similar to those used in Explorer-17 [48] and in rocket experiments with thermophysical probes [8] were published by Brace and Reddy [49]. Unlike the previously available data, obtained from other satellites which reached lower latitudes, Explorer-22 revealed not a monotonic increase of T_e with increasing latitude, but the presence of clear-cut latitudinal maxima of T_e with increasing southern geomagnetic latitude above ~60° and the northern latitude above ~40°; at an altitude of ~1000 km, T_e does not increase but decreases (Fig. 14).

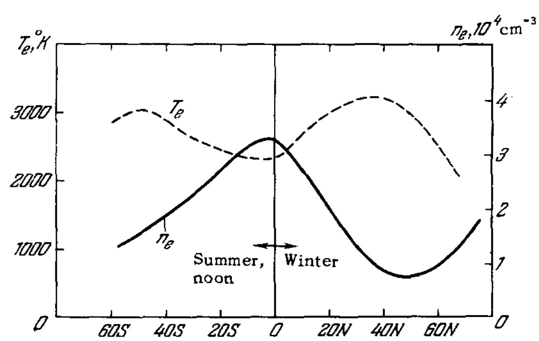

FIG. 14. Latitudinal variations of T_e and n_e, measured on the satellite Explorer-22 [49].

Figure 15 shows the diurnal variation of T_e (from the data of Explorer-22 - upper solid curve and points [49], and from the data of Ariel-1 - lower solid curve [43]) and of n_e (from the data of Explorer-22 [49]). The plots pertain to an altitude ~1000 km and geomagnetic latitude 40°. The daytime values of T_e, determined from measurements on Explorer-22, exceed the corresponding values of T_e as given by Ariel-1 by approximately 1000°. The low values of T_e measured by Ariel-1 are difficult to explain, since in 1962 the values of S were higher than in 1964, and according to [46] T_e increases with increasing S.

One of the possible explanations of this discrepancy between the results is the decrease of n_e, which apparently takes place in the ionosphere at all altitudes with decreasing solar activity,* and the associated deterioration of the electron-cooling conditions, which causes an increase in T_e. Another possible cause may be connected with the peculiarities of the reduction (particularly, averaging) of the data of Ariel-1, which are difficult to assess, since the "individual" values of the data corresponding to

*This is particularly corroborated by comparison of the n_i data obtained with the third Soviet satellite and with Kosmos-2 (see [50]).

each individual orbit of Ariel-1, as already noted, have not been published.

In all the mentioned satellites, the temperatures of the charged particles were determined by different modifications of probes. Some information on T_e and T_i can be obtained, however, by processing ionograms obtained by pulsed radio sounding of the ionosphere operated with the aid of ionospheric stations mounted on the satellites (similar to Alouette).

As is well known, from the data obtained by sounding the atmosphere upward, starting with the maximum of the F region, the value of n_e decreases with altitude. If we approximate $n_e(H)$ by an exponential function

$$n_e = n_{e0}e^{-H/\overline{H}_e},$$

then $\overline{H}_e$ can be called the plasma altitude scale.

It can be shown (Watt[51]) that in a neutral ionosphere which is in diffusion equilibrium we have

$$\overline{H}_e = \frac{T_e + T_i}{\frac{\partial T_e}{\partial H} + \frac{\partial T_i}{\partial H} + \frac{\langle \overline{m}_i \rangle g}{k}},$$

where $(\overline{m}_i)$ is the average ion mass, g the gravitational acceleration, and k Boltzmann's constant.

We see from this expression that $\overline{H}_e$ is determined by five parameters of the ionosphere: T_e, T_i, $\partial T_e/\partial H$, $\partial T_i/\partial H$, and $\langle m_i \rangle$. It is therefore difficult to obtain any of the parameters by experimentally determining $\overline{H}_e$ *only.*

Watt[51] determined $\overline{H}_e$ from ionograms obtained with the aid of Alouette-1 for the altitude region from 400 to 900 km and for geomagnetic latitudes from 48°N to 78°N. He used a number of assumptions simplifying the structure of the ionosphere and of the processes occurring in it (in particular, he assumed *no horizontal gradients of* n_e, diffusion equilibrium in the entire interval of altitudes under consideration, thermal equilibrium between ions of different masses). Some of the other assumptions were based on Hanson's theoretical considerations (it was assumed that $T_e = T_i$ at $H = 800$ km) and on non-simultaneous rocket measurements of the ion-mass spectrum by Taylor et al.[52]. On the basis of such assumptions, Watt calculated the values of T_e and T_i as functions of the latitude in the altitude interval 500–800 km for daytime and nighttime conditions of the summer of 1963 and the Winter of 1963–1964. The latitude distributions of T_e and T_i obtained in[51] have maxima in the region of high latitudes, recalling the maxima observed as the results of measurements of T_i on the Explorer-22.

Although the ionograms obtained when sounding the ionosphere upward undoubtedly contain information concerning the temperatures of the charged particles, and a data reduction similar to that carried out by Watt[51] is advisable, it must be borne in mind that the accuracy and reliability of the values of T_e and T_i obtained by such a method are much lower than when these ionograms are used to determine n_i or when T_e and T_i are measured by probe methods.

The only experiments in which T_e were determined in the peripheral region of the ionosphere (much higher than 1000 km) were carried out by Serbu and Maier with the IMP-2 satellite with the aid of a three-electrode flat charged-particle trap by the retarding-potential method[53]. According to their conclusions, which are based on an analysis of data obtained during one-half year, T_e increases with altitude like $\sim R^2$ (where R is the geocentric distance), and n_e decreases like $\sim 1/R^3$ down to $R \sim 5R_E$ (R_E - earth's radius). From $R \sim 5R_E$ to the apogee, which is equal to 15.9 R_E, T_e varies little, amounting to ~1–2 eV, and n_e also changes little.

These are very interesting and impressive results; apparently, so far only a slight fraction of all the data available to the authors has been reported. We wish to note, however, that the results of[53] raise certain questions which will probably be clarifed in later publications. Thus, it is not quite clear with which accuracy the T_e was determined from the probe characteristic in the case when T_e is 0.1–0.2 eV, since the retarding potential varies, assuming discrete values spaced ~1 V apart. It is somewhat difficult to explain the absence from all the altitude distributions of n_e obtained in[53] of the "knee" - the accelerated decrease of n_e (or, equivalently, of n_i) near $R \sim (4–5)R_E$, which was observed many times by different methods and by three independent groups of observers[36,20,21]. Regardless of these remarks (which can turn out to be insignificant in light of further publications), however, the very fact that an appreciable growth of T_e *with increasing* altitude has been observed in the peripheral region of the ionosphere is very interesting.

It should be noted that no theoretical calculations were made of $T_e(H)$ at altitudes on the order of several times R_E; the only theoretical work pertaining to this problem is the already mentioned article by Geisler and Bowhill[19], who considered the variation, along a geomagnetic tube of the temperature of an ionospheric plasma heated by fast photoelectrons rising along the tube from the F region; a calculation made for a geomagnetic tube crossing the 1000 km level at geomagnetic latitude ~40° yielded an almost isothermal distribution of T_e along the tube (up to the tube length ~8000 km) with temperatures ~3000°K (under conditions of minimum solar activity). This calculation does not contradict the measurement data reported in[53], but does not predict the growth of T_e with altitude observed in[53].

Of course, the photoelectrons with honest energy can rise along the geomagnetic force lines with small energy losses, *since their cross section* q_i *for interaction with the ions decreases approximately like*

v_e^{-4} (where v_e is the electron velocity), and possibly it is they which cause the high temperatures of the plasma in the peripheral part of the ionosphere. One cannot exclude, however, the existence of some thermal connection between this highest part of the ionosphere and the solar-wind plasma, the energy of which penetrates into the magnetosphere by some mechanism (for example, that proposed by Dessler and Walter [54] or considered by Block [55]).

It is obvious that to establish the source of heating of the peripheral region of the ionosphere it will be necessary to perform new investigations (both experimental and theoretical).

V. CONCLUSION

In the beginning of this review we already noted that it makes no reference at all of the numerous and therefore exceedingly valuable temperature data obtained as a result of land-based observations (in particular, those reported in [56,57]). This was done not only to be able to avoid as much as possible duplication with Evans'es review [1], but also because, in the author's opinion, it is interesting to see what conclusions can be drawn regarding the particle temperatures in the ionosphere by using only the results of experiments carried out with the aid of instruments carried by rockets and satellites.

These conclusions turn out to be as follows:

The most reliable are direct measurements of T_g carried out with the aid of mass spectrometers and with the aid of observations of fluorescence of K, Na, and AlO. These measurements are not plentiful. Almost all the measured values of T_g in the altitude interval from 100 km to ~200 km increase with altitude. The dependence of T_g on the intensity of the solar short-wave radiation (characterized by the flux of decimeter radiation S) apparently increases with increasing altitude (to ~300 km). The values of T_g measured by direct methods agree quite satisfactorily with theoretical models based on the results of observation of satellite deceleration.

Many of the considered results of the measurements of T_g either point to the existence of an isothermal zone at altitudes H ~ 200—300 km (the mass-spectrometric experiments of 1961 [23] and of 1963—1964 [8]), or else do not contradict its existence (experiments on measuring the broadening of the resonance lines of K, Na, and AlO [25 26]) and point to a dependence of the altitude and of the temperature of this zone on S and on the local time. Nonetheless, these data are not plentiful enough to state that such a zone always exists.

There is a communication reporting deviations from isothermy at these altitudes (possibly of short duration and therefore not revealed as a result of observations averaged over prolonged periods [28]). The values of T_g at the heights of the isothermal zone, determined with the aeronomic satellite Explorer-17, are characterized by a considerable scatter, and the results of the final analysis of the causes of this scatter have not yet been published.

Measurements of T_i on satellites and rockets are quite few. Daytime values of T_i are much lower than the values of T_e at altitudes 200 km < H < 400 km [32,34] and at altitudes 400—600 km [33], where the presence of a considerable scatter of T_i from day to day (amounting to several dozen of hundreds of degrees) has been established; there is a tendency for T_i to increase with increasing latitude [33]. At altitudes H ~ 5000—8000 km, T_i is in any case lower than 9000—10 000° [35]. No direct simultaneous measurements of T_i and T_g were made.

Measurements of T_e on rockets and satellites are much fewer in number than measurements of T_g and T_i. There is no thermal equilibrium at altitudes 200 km in the daytime ionosphere ($T_e > T_i > T_g$) ([7,45,9,10] and elsewhere). In the nighttime ionosphere, there is likewise no thermal equilibrium [46,48]. In 1961, daytime T_e(H) profiles were obtained with maxima of T_e at altitudes ~220 km [7]; these were close in form to the theoretical models of [13] and [14]. During 1962 and later, most daytime measurements at altitude H ~ 220 km revealed an increase of T_e with altitude [34,41,43] (although one T_e(H) profile was obtained in 1963 with a maximum of T_e at H ~ 250 km [7]). It is possible that the form of the T_e(H) profile at altitudes H ≤ 1000 km varies with the phase of the solar-activity cycle, as follows from theoretical calculations [16]; a final confirmation of this can be obtained by measuring T_e(H) during the time of the next solar-activity maximum.

Rocket measurements of T_e(H) during the time of the solar eclipse on 20 July 1963 (and simultaneous rocket measurements of the intensity of the short-wave solar radiation in different sections of the spectrum) offer convincing evidence that the main source of heating of electrons in the daytime F region is solar radiation, since the electrons in the E region, which is ionized by x rays from the sun, have a different main source of heating (possibly, the electric field) [37,38].

The existence of a peak in the diurnal variation of T_e has been established near sunrise [45,46,47]; this peak decreases with increasing altitude [46].

In most measurements, an inverse correlation between T_e and n_e is observed [46,49].

Magnetic perturbations lead to an increase of T_e according to the rocket measurements [7], and to a decrease according to measurements on the Ariel-1 satellite [46]. The published results of measurements with rockets and satellites do not make it possible to draw clear-cut conclusions concerning the influence of magnetic disturbances on T_e.

The daytime distribution of T_e over the geomagnetic latitudes has a maximum near the equator at an altitude of ~1000 km, and maxima near L ~ 40°N and L ~ 50°S. During the nighttime, these maxima shift towards higher geomagnetic latitudes[49].

It has been observed that in the peripheral part of the ionosphere (at H ≫ 1000 km), T_e increases with altitude in proportion to R^2 (R - geocentric distance), rising to 10 000—20 000° at altitudes ~25 000 km[53]. No theoretical models for this region have been published as yet.

[1] J. V. Evans, The Temperature of Neutral and Charged Particles in the Ionosphere, Paper delivered at Symposium on Physics of Solar-Terrestrial Relations, Belgrade, August, 1966 (Preprint).

[2] R. E. Bourdeau, Space Res. 5, 77 (1965).

[3] J. V. Evans, J. Geophys. Res. 70, 17, 4365 (1965).

[4] T. K. Breus and G. L. Gladevich, Kosmicheskie issledovaniya (Cosmic Research) 3, 877 (1965).

[5] K. S. W. Champion, Atmospheric Structure in the lower thermosphere. Air Force Cambridge Research Labs, Surveys in geophysics. N168, 1965.

[6] Y. Aono, K. Hirao and S. Miyasaki, J. of the Radio Research Labs 46, 407 (1962).

[7] L. H. Brace, N. M. Spencer and G. R. Carignan, J. Geoph. Res. 68, 5397 (1963).

[8] L. H. Brace, N. M. Spencer, G. R. Carignan, D. R. Taeusch and N. Niemann, J. Geoph. Res. 70, 2665 (1965).

[9] K. I. Gringauz, B. N. Gorozhankin, G. L. Gdalevich, N. M. Shutte, R. E. Rybchinsky and V. V. Afonin, Space Res. 5, 773 (1965).

[10] B. P. Bowan, R. L. F. Boyd, G. L. Henderson and A. P. Willmore, Proc. Roy. Soc. 281, 514 (1964).

[11] I. Harris and W. Priester, Space Res. 3, 53 (1963).

[12] M. N. Izakov, in: Issledovaniya kosmicheskogo prostranstva (Outer Space Research), Nauka, 1965, p. 30.

[13] W. B. Hanson and F. S. Johnson, Memories Soc. Roy. Sc. Liege 4, 390 (1961).

[14] W. B. Hanson, Space Res. 3, 282 (1963).

[15] A. Dalgarno, N. B. McElroy and R. J. Moffet, Planet. Space Sci. 11, 463 (1963).

[16] J. E. Geisler and S. A. Bowhill, J. Atmosph. Terr. Phys. 27, 457 (1965).

[17] K. I. Gringauz, V. V. Bezrukhin, V. D. Ozerov and R. E. Rybchinskiĭ, Dokl. Akad. Nauk SSSR 131, 1301 (1960) [Sov. Phys.-Dokl. 5, 361 (1960)].

[18] K. I. Gringauz, V. G. Kurt, V. I. Morozov and I. S. Shklovskiĭ, Astron. Zh. 37, 716 (1960) [Sov. Astron.-AJ 4, 680 (1961).

[19] J. E. Geisler and S. A. Bowhill, Atmosph. Terr. Phys. 27, 1119 (1965).

[20] B. A. Taylor, Jr., H. C. Brinton and C. R. Smith, J. Geoph. Res. 70, 5769 (1965).

[21] D. L. Garpenter, J. Geoph. Res. 71, 693 (1965).

[22] A. A. Pokhunkov, in: Iskusstvennye sputniki Zemli (Artificial Earth Satellites) 13, 110 (1962).

[23] A. A. Pokhunkov, in: Ionosphernye issledovaniya (Ionospheric Research) 14, 26 (1965).

[24] J. E. Blamont, T. M. Donahue and M. L. Lory, Phys. Rev. Letts 6, 403 (1961).

[25] J. E. Blamont and M. L. Lory, Space Res. 5, 1137 (1965).

[26] J. E. Blamont and M. L. Lory, A paper presented at the 6-th COSPAR Symposium, 1965.

[27] N. G. Spencer, L. H. Brace, G. R. Carighan, D. R. Taousch and H. Niemann, J. Geoph. Res. 70, 2665 (1965).

[28] V. V. Mikhnevich, in: Issledovaniya kosmicheskogo prostranstva (Outer Space Research), 1965.

[29] G. P. Newton, R. Horowitz and W. Priester, Planet. Space Sci. 13, 599 (1965).

[30] A. Ya. Pressman and S. P. Yatsenko, Geomagnetizm i aéronomiya (Geomagnetism and Aeronomy) 1, 49 (1961).

[31] G. W. Sharp, W. B. Hanson and D. D. McKibbin, Space Res. 4, 454 (1964).

[32] V. V. Afonin, T. K. Breus, G. L. Gdalevich, B. N. Gorozhanskii, K. I. Gringauz, R. E. Rybchinskiĭ, and N. M. Shyutte, in: Issledovaniya kosmicheskogo prostranstva (Outer Space Research), Nauka, 1965, p. 151.

[33] R. L. F. Boyd and W. J. Rait, Space Res. 5, 207 (1965).

[34] A. F. Nagy, L. H. Brace, G. R. Garignan and M. Kanal, J. Geoph. Res. 68, 6401 (1963).

[35] K. I. Gringauz, V. V. Bezrukikh and T. K. Breus, Kosmicheskiye issledovaniya (Cosmic Research) 5, 245 (1967).

[36] K. I. Gringauz and V. V. Bezrukikh, in: Issledovaniya kosmicheskogo prostranstva (Outer Space Research), Nauka, 1965, p. 77.

[37] Y. Aono, K. Hirao, S. Miyasaki, K. Takayama and T. Dote, Space Res. 2, 1215 (1961).

[38] L. G. Smith, C. A. Accardo, L. H. Weeks and P. J. McKinnon, J. Atmosph. Terr. Phys. 27, 803 (1965).

[39] N. D. Papaleksi, in: Trudy ekspeditsii po nablyudeniyu polnogo solnechnogo zatmeniya 19 iyulya 1936 g. (Transactions of Expedition to Observe the Total Solar Eclipse of 19 July 1936), AN SSSR, 1938, p. 115.

[40] J. C. Ulwick, W. Pfister and R. E. McInerney, Space Res. 5, 171 (1965).

[41] K. I. Gringauz, G. L. Gdalevich, V. F. Gubsky, I. A. Knorin, V. A. Rudakov and N. M. Shutte, A paper presented at the 7-th COSPAR Symposium in Vienna, 1966.

[42] P. J. Bowen, R. L. F. Boyd, G. L. Henderson and A. P. Willmore, Proc. Roy. Soc. 281, 526 (1964).

[43] K. Hirao, Presented at the 7-th COSPAR Symposium in Vienna, 1966.

[44] R. E. Bourdeau, Space Res. **2**, 554 (1961).
[45] R. E. Bourdeau and J. L. Donley, Proc. Roy. Soc. **281**, 1387 (1964).
[46] A. P. Willmore, Proc. Roy. Soc. **286**, 537 (1965).
[47] R. C. Sagalyn, M. Smiddy and Y. N. Bhargaya, Space Res. **5**, 189 (1965).
[48] B. H. Brace, N. W. Spencer and A. Dalgarne, Planet, Space Sci. **13**, 647 (1965).
[49] L. H. Brace and B. M. Reddy, J. Geoph. Res. **70**, 5783 (1965).
[50] K. I. Gringauz, B. N. Gorozhankin, N. M. Shutte and G. L. Gdalevich, Space Res. **4**, 473 (1964).
[51] T. M. Watt, J. Geoph. Res. **70**, 5849 (1965).
[52] H. A. Taylor, L. H. Brace, H. C. Brinton and C. R. Smith, J. Geoph. Res. **68**, 5339 (1963).
[53] G. P. Serbu and E. J. Maier, J. Geoph. Res. **71**, N 15 (1966).
[54] A. J. Dessler and G. K. Walter, Planet. Space Sci. **12**, 227 (1964).
[55] L. P. Block, J. Geoph. Res. **70**, 885 (1965).
[56] J. V. Evans, Planet. Space Sci. **13**, 1031 (1965).
[57] J. V. Evans, J. Geoph. Res. **70**, 2726 (1965).
[58] A. I. Repnev, Trudy, TsAO (Cent. Astron. Obs.), No. 25, 5 (1959).

Translated by J. G. Adashko