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## Chapter X

# Rocket and Satellite Measurements of Ionospheric and Magnetospheric Particle Temperatures

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## X.1. Introduction

In this review of the temperatures of neutral particles ($T_g$), ions ($T_i$) and electrons ($T_e$) in the earth's ionosphere, only data obtained by means of rocket- and satellite-borne devices are used, since the results derived from ground-based observations (including the determination of $T_g$ from measurements of satellite drag) have been reviewed by J. V. Evans in Chapter IX.

The compilation of this paper was considerably facilitated by the fact that in recent years several good reviews have been published concerning temperature measurements in the ionosphere; these include those of Bourdeau (1965), Evans (1965b), Breus and Gdalevich (1965), and a section on temperatures in a report by Champion (1965). Nevertheless, additional experimental data were published during 1965 and 1966, and these should be analysed and compared.

The temperatures of the neutral and charged particles are among the most important properties of the ionosphere. By comparing the distributions of

$T_g$, $T_i$ and $T_e$ one can estimate and localize the known heat sources and establish the existence of other heat sources not yet detected by present-day instruments. It is thus expected that increasing attention will be devoted to ionospheric temperature investigations.

Although consideration of the techniques of determining $T_g$, $T_i$ and $T_e$ lies outside the scope of the present paper it seems reasonable to remark on the role and some specific features of temperature measurement performed by means of instruments flown on rockets and satellites.

For the determination of ionospheric temperatures ground-based measurements (as well as other non-rocket investigations) have considerable advantages over those using rockets, as they enable statistically extensive data to be obtained at comparatively low cost. Ground-based observations of satellite drag have served as a basis for modern models of the upper atmosphere, including temperature models. The development of ionospheric investigations using the incoherent back-scatter method (considered in Chapter IX) has enabled some ionospheric parameters to be determined with reasonable confidence in their validity.

Nevertheless, such measurements have some disadvantages which are difficult to eliminate. For instance, in the results obtained from satellite drag observations it is impossible, without the introduction of additional data, to separate the effects caused by a change in mass composition from those produced by the temperature variation. The method is characterized by some ambiguity and by poor time and height resolution. Such peculiarities are, to some extent, also typical of the method of ionospheric investigation by means of incoherent back-scatter (see Evans, 1965b).

By using instruments flown in rockets it is feasible, in principle, to determine the parameters of the medium unambiguously and with high resolution in time as well as in height. Therefore, although ionospheric rocket launchings are relatively rare on account of the high cost, they are of paramount significance since a comparison of the results of rocket measurements with those of ground-based measurements of the same parameters raises confidence in the correctness of the ground-based measurements. In some cases rocket methods are capable of providing information which is unobtainable by any other method.

Temperature measurements from rockets and satellites, coupled with the detection and analysis of low-energy particles, are among the most difficult of rocket experiments. Therefore most of the experimenters have endeavoured to ensure the verification of measured results. For instance, in Japanese rockets several types of probe were used for the simultaneous measurements of $T_e$ (Aono *et al.*, 1962). During the launching of some rockets in the U.S.A. measurements were carried out on probes separated from the rockets (in order to avoid the disturbance of the medium by the rocket) (Brace *et al.*, 1963; Spencer *et al.*, 1965). Simultaneously, use was made of identical probes



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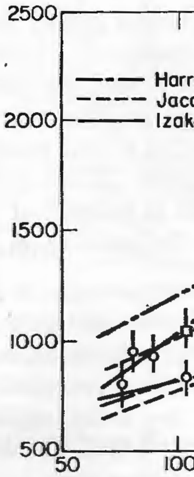


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located in different positions on a satellite. This was done on the Soviet satellite Cosmos 2 (Gringauz *et al.*, 1965) and on the U.S.A.–U.K. satellite satellite Ariel 1 (Bowen *et al.*, 1964a). In the U.S.A. a rocket was launched to pass comparatively closely to a satellite (Evans, 1965b). Data obtained by means of rocket probes were compared with those of ground-based measurements (Spencer *et al.*, 1965).

The experimental findings have led to the conclusion that, in most cases, data on particle temperatures obtained by means of rocket and satellite-borne instruments are sufficiently reliable.

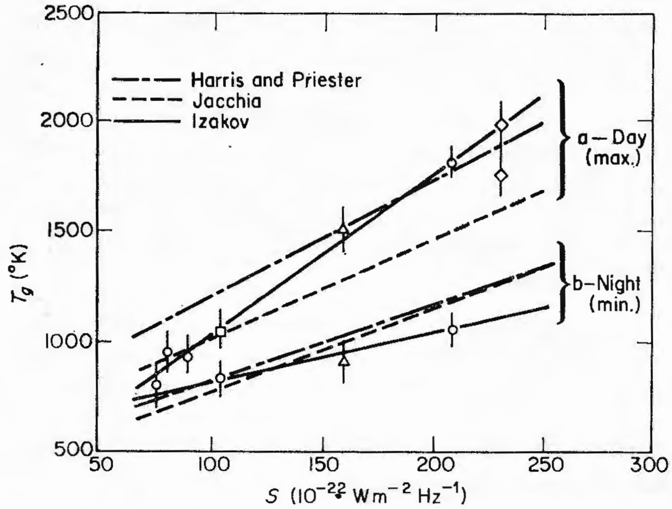


FIG. X.1. The variation of neutral gas temperature, $T_g$, in the isothermal zone with solar radiation flux $S$. (After Izakov, 1965.)

The main rocket and satellite experiments concerning particle temperature distribution in the ionosphere were performed after 1960. The basic theoretical and semi-empirical models were also developed in the sixties.

According to the model proposed by Harris and Priester (1963), $T_g$ increases monotonically to a height where a "thermopause" is situated which forms the base of the isothermal zone. The height of the thermopause and the value of $T_g$ in the isothermal zone vary as a function of the $S$ index (that is the flux of 10·7 cm solar radiation in $10^{-22}$ Wm$^{-2}$Hz$^{-1}$) and local time.

The task of correlating the models of the upper atmosphere and experimental data continues. Figure X.1 from one of the more recent papers (Izakov, 1965) shows $T_g$ in the isothermal zone as a function of $S$ according to Harris and Priester, Jacchia and Izakov.

Calculations of the value of $T_e$ appropriate to the daytime ionosphere were performed by Hanson and Johnson (1961), Hanson (1963), Dalgarno *et al.*

(1963) and, more recently, by Geisler and Bowhill (1965a). The calculations of Hanson and of Dalgarno *et al.* were published in 1963 when detailed profiles of $T_e$ as a function of altitude had not yet been obtained experimentally, and they are based on the following scheme:

(a) firstly, the rate of photo-ionization, and the processes of energy loss by photoelectrons are considered and the rate of heat inflow per unit volume of electron gas in the ionosphere is thus determined;

(b) next, the rate of cooling of the electron gas by means of inelastic collisions with neutral particles below 250 km, and by elastic (Coulomb) collisions with ions above 250 km is considered. Hanson (1963) has taken into account the thermal conductivity of the electron gas at heights greater than 600 km, whereas Dalgarno *et al.* have ignored thermal conductivity. In both models (Hanson, 1963; Dalgarno *et al.*, 1963) the rates of heating and cooling of the electron gas are assumed to be equal at all altitudes, and this makes it possible to determine the height distribution of $T_e$. According to both models $T_e$ starts to increase relative to $T_g$ above a height of about 120 km and reaches a maximum at about 220 km; thermal equilibrium is restored at a height of about 350 km. According to Hanson and also Dalgarno *et al.* at a height $H \gtrsim 300$ km the equation

$$T_e - T_g = 2 \times 10^6 \frac{Q}{N_e^2} T_e^{3/2}$$

is valid, where

$Q$ = heat flux per cm³ of electron gas (in eV cm$^{-3}$ sec$^{-1}$);

$N_e$ = electron density per cm³.

As far as $T_i$ is concerned, Hanson (1963) has concluded that, when $H \gtrsim 1000$ km, $T_i \to T_e$.

Bourdeau (1965) pointed out the need for theoretical consideration of the possibility of non-local dissipation of solar ultraviolet radiation energy, that is, the possibility that absorbed energy may be transferred to other heights. Geisler and Bowhill (1965a) showed in their calculations that it is necessary, especially under sunspot-minimum conditions, to take into account the thermal conductivity of the electron gas at much lower heights than was done by Hanson (1963). If thermal conductivity is taken into account, i.e. the possibility of heat transfer to higher altitudes is considered, $T_e(H)$ profiles considerably different from those calculated by Hanson and Dalgarno *et al.* are obtained; not only does $T_e$ not decrease with height above 220 km, but $T_e$ may even have a finite positive gradient. Geisler and Bowhill (1965a) conclude that, under conditions of solar maximum, the $T_e(H)$ profiles would have a form close to that calculated by Hanson and by Dalgarno *et al.* The problem of the agreement of experimental $T_e(H)$ profiles with theoretical ones in case of moderate or low solar activity will be dealt with in Section X.4.

As far as the $T_e(H)$ profiles at sunspot maximum are concerned, we must await the next solar maximum before full experimental data will be available for comparison with theoretical models because, as was pointed out above, no such experimental data exist for 1957 and 1958.

Since 1959, when charged-particle experiments on Soviet spacecraft produced reliable evidence about the earth's plasma envelope at heights of up to 20 000 km approximately (Gringauz *et al.*, 1960a), there has not been a single agreed name for this region of space. For example, it has been variously called an "ionized component of the geocorona" (Gringauz *et al.*, 1960b), the "protonosphere" (e.g. by Geisler and Bowhill, 1965a, b), the " magneto-ionosphere" (Taylor *et al.*, 1965), and the "plasmasphere" (Carpenter, 1966), etc. It seems that the term "outermost region of the ionosphere" or "outermost ionosphere" may be the most appropriate for this zone; this term will be used throughout the present paper.

On the basis of the theoretical estimates of Geisler and Bowhill (1965b) $T_i = T_e = T$, for altitudes much higher than 1000 km; this is also in accordance with Hanson (1963). According to Geisler and Bowhill (1965b) $T$ should vary between 3000° and 3400°, depending on the phase of the solar cycle, at a distance of 8000 km along a magnetic line of force crossing the 1000 km height level at a geomagnetic latitude of 40°.

In concluding this section, we should note that all the theoretical calculations of charged-particle temperatures in the ionosphere (Hanson and Johnson, 1961; Dalgarno *et al.*, 1963; Hanson, 1963; Geisler and Bowhill, 1965a) comprise stages where not entirely reliable estimates have been used; one problem, for instance, is the determination of the production rate of photoelectrons and their average energies at different heights. Further, there is some doubt about possible ways of heating charged particles (for example, non-thermal particle fluxes, magnetohydrodynamic waves, electric fields) which have not been taken into account in the theoretical calculations. It would therefore be unreasonable to expect complete agreement between experimental results and calculations.

## X.2. Measurements of Neutral-Particle Temperature, $T_g$

The determination of the neutral-particle temperature in the upper atmosphere is mainly performed by one of two methods. In the first of these, after a preliminary determination of the height density distribution, $T_g$ is computed from the scale height. The other method is based on the ejection of Na, K or AlO from a rocket into the region under investigation, followed by the measurement of the Doppler broadening of the resonance lines (or in the case of AlO, the bands).

Both methods yield data averaged over a height range. In the second case the averaging interval is governed by the dimensions of the luminescent cloud; furthermore, the luminescence observed on the earth is determined not only by the particles at the surface of the cloud, but also by those along the entire line of sight to the observer.

The scale height $H = RT_g/mg$ essentially depends on the average molecular weight $m$. For an accurate determination of $T_g$ it is, therefore, necessary to know the mass spectrum of particles in the region under investigation. If $H$ is found by mass-spectrometer measurements for one component of a neutral gas mixture, the value of $T_g$ may be determined with higher accuracy than from data on the atmospheric density variations with height, obtained by means of rocket-borne manometers (not to mention data based on the analysis of satellite drag).

This higher accuracy is due to the fact that, in mass-spectrometer experiments there is no need to make assumptions regarding the mean molecular weight of particles, whereas such assumptions are necessary if the manometric method is used in the absence of a mass-spectrometer.

In Fig. X.2 the solid curves represent the values of $T_g$ derived from Soviet mass-spectrometer measurements. These results were obtained during launchings of geophysical rockets at middle latitudes in the U.S.S.R. (Pokhunkov, 1962 and 1965). The value of $T_g$ was determined by two methods: (a) by means of measurements of the changes with height in the relative concentration of two inert gases, and (b) by measurements of the height distribution of partial pressure for one component of a mixture of neutral gases. To enable the temperature to be measured by method (a) there must be stable gravitational separation, which rocket experiments have confirmed to exist at heights above 110 km. The errors in the determination of $T_g$ are estimated to amount to 10% of the measured values. In all three experiments the increase of $T_g$ is observed to begin at about 100 km. The first two measurements correspond to an average level of solar 10·7 cm flux, $S$, of 175 (in units of $10^{-22}$ W m$^{-2}$ Hz$^{-1}$). During the third measurement (1961) $S$ equalled 100. According to these data, the thermopause (the lower boundary of the isothermal zone) lies much higher than 200 km (during the experiment of 15 November, 1961, at a height above 300 km, with $T_g$ in the isothermal zone being about 1 500°). It should be noted that the existence of the isothermal zone can be detected only on the curve of 15 November, 1961. Apparently the first two measurements were not made at sufficiently great heights.

Blamont *et al.* (1961) have reported on a number of experiments aimed at determining $T_g$ by the ejection of chemical reagents from rockets launched during morning and evening twilight conditions in Algeria and in the U.S.A. The broadening of the K and Na resonance lines and the AlO bands fluorescing under the influence of solar radiation was studied. Figure X.2 also shows some values of $T_g$ determined by Authier *et al.* (1965) and Blamont and Chanin-Lory (1965). The number of measurements carried out by Blamont's group at heights above 200 km is unfortunately insufficient to prove the

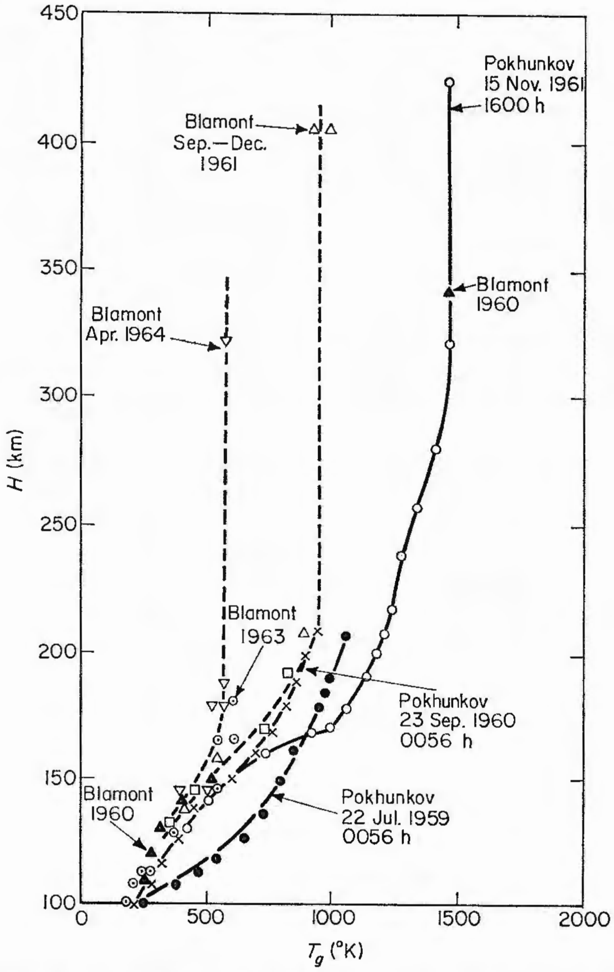


**FIG. X.2.** Values of the neutral gas temperature measured at different heights by Pokhunkov and by Blamont *et al.*

existence of the isothermal zone and to determine its temperature and the height of the thermopause. No measurements were obtained for heights between about 210 and 410 km in Blamont's 1961 experiment and, had there been a $T_g$ minimum or maximum in this region, it would not have been detected. All the data relating to heights below about 150 km and obtained in different years by the various method lie comparatively close together and appear to indicate that $T_g$ at these heights depends very little on the solar cycle.

In their determination of $T_g$, Spencer *et al.* (1965) employed devices which were separate from the four rockets launched between 1962 and 1964 above Wallops Island (Virginia, U.S.A.) at different times during the day, at sunset, at night and also during the solar eclipse of 20 July 1963. These devices were called "thermospheric probes", and they contained, apart from Langmuir probes, an omegatron mass-spectrometer arranged to determine the partial concentration of one component of the neutral atmosphere. These very interesting experiments have demonstrated the close agreement of experimentally determined values of $T_g = T_{N_2}$ with those derived from the model of Harris and Priester (1963). Figure X.3 shows various $T_g(H)$ profiles obtained during these four experiments. (The $T_e$ data obtained during the same experiments will be presented in Section X.4.) In all four $T_g(H)$ profiles the existence of the isothermal zone is clearly evident. The height of the thermopause for a local time of about 16 h appears to be lower

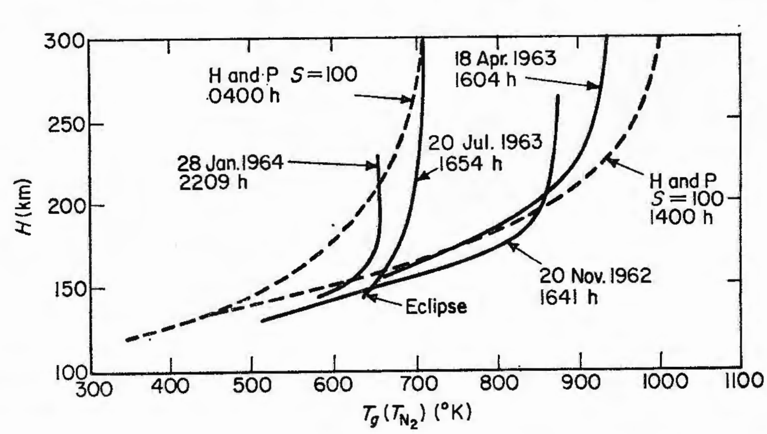


FIG. X.3. Values of $T_g(H)$ determined by Spencer *et al.* (1965) in four rocket firings during 1963/64.

than that given in the 1961 results of Fig. X.2. This may be interpreted as evidence of a smaller heat inflow into the upper atmosphere in 1963 than 1961. For the same reason the height of the thermopause is lower near sunset than midday.

Although, from a statistical point of view, the results of rocket measurements obtained by Pokhunkov, Blamont, Spencer, Brace *et al.* are certainly not comparable with the results of $T_g$ determinations based on satellite drag observations, they are, nevertheless, very important because they yield high resolution in both time and altitude, and they can be interpreted without ambiguity.

The results of the above-mentioned rocket measurements appear to confirm the existence of the isothermal zone above 200–300 km; at least the results do not disprove that such a zone exists. However, according to some published results, irregular variations of $T_g$ were observed at the above-mentioned heights. For example, in analysing the results of manometric measurements made on geophysical rockets reaching 450 km, Mikhnevich (1965) concluded that minima of $T_g$ existed at heights between 200 and 300 km.

It seems that the data on $T_g$ $(H)$ profiles currently available are not sufficient to disprove that height distributions similar to those found by Mikhnevich may exist for short periods at least.

To determine $T_g$, use has also been made of instruments aboard the Explorer 17 aeronomic satellite placed in April 1963 into an eccentric orbit with a perigee of 258 km and an apogee of 420 km. Newton *et al.* (1965) determined $T_g$ from the variations with height of the density measured by manometers on this satellite. The average mass of the gas particles was derived from data obtained by means of a mass-spectrometer. The experimenters estimate the

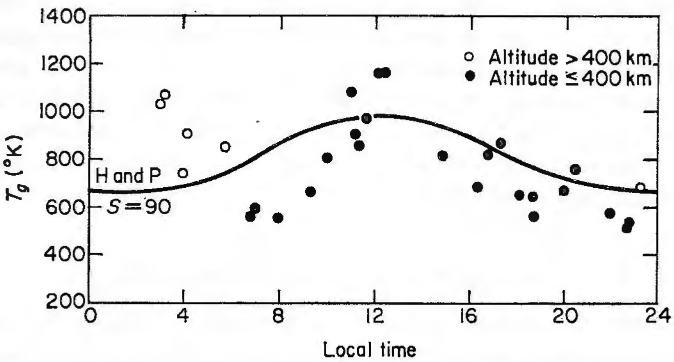


**Fig. X.4.** $T_g$ values determined by manometers on Explorer 17. The solid curve shows the Harris and Priester values for $S$=90. (After Newton *et al.*, 1965.)

error in the determination of $T_g$ to be less than 20%. Figure X.4 shows the $T_g$ values obtained at different local times; the solid curve indicates the dependence of $T_g$ on local time given by the Harris and Priester model for $S$=90. The $T_g$ values reported by Newton *et al.* vary by a factor of 2 approximately (from about 500° to about 1 000°). The considerable dispersion of the points during the period 0300 to 0700 L.M.T. is evidence of a large dispersion of scale heights. Newton *et al.* state that the reasons for this are not clear, but may, perhaps, be accounted for by variations in temperature, or in the average mass of the particles, or both simultaneously, or else by a departure from diffusive equilibrium. Newton *et al.* (1965) have indicated that the analysis of results obtained in the height range 500–600 km will be continued, particularly in order to separate the effects of composition and temperature changes.

All the $T_g$ data described above were obtained from measurements of physical characteristics (scale height, Doppler broadening of resonance lines) observed by averaging over comparatively large ionospheric regions. It is feasible in principle, however, to make direct absolute $T_g$ measurements, which are localized to a much greater extent. This feasibility was pointed out by Pressman and Yatsenko (1961) who suggested that use should be made of measurements of particle flux entering two narrow tubes which are positioned at different angles to the velocity vector of a space vehicle; the appropriate theory was given. The difficulties in the implementation of such measurements lie purely in engineering and are connected with the problems of (a) determining accurately the orientation of a space vehicle, and (b) the measurement of very small currents. Similar difficulties are involved in the "velocity scan method" suggested by Spencer *et al.* (1965); in this method $T_g$ is determined by studying the depth of modulation of the current from the omegatron mass-spectrometer produced during a periodic change in the orientation of its input hole as the thermospheric probe rotates. Preliminary results of such absolute measurements of $T_g$ based on mass-spectrometer current modulation were published, but appear to be somewhat overrated. However, such methods are undoubtedly very promising, and there is every reason to believe that the technical difficulties will be overcome and the reliability and, particularly, the height resolution in the determination of $T_g$ will be considerably improved.

## X.3. Measurements of Ion Temperature, $T_i$

The number of successful attempts to measure $T_i$ directly by means of rocket- and satellite-borne instruments, and those which are described in the literature, is not large.

Sharp *et al.* (1964) reported an experiment aimed at the direct determination of $T_i$ by measuring the ion velocity distribution by means of a planar ion trap using a method of retarding potentials. Such traps were carried by two satellites launched in 1961 and 1962 into near-circular high-inclination orbits. These traps were of plane geometry and were mounted facing the direction of travel of the satellites, i.e. they pointed very closely along the velocity vector. For the first satellite (at heights between about 230 and 240 km) the calculated values of $T_i$ oscillated between about 1 200° and about 2 400° from one cycle of retarding potentials to another. With the second satellite (at heights from 245 to 280 km) $T_i$ oscillated from about 600° to about 1 800°; in some cases the $T_i$ values were found to be lower than the expected $T_g$ values. The authors state that, since the instrument should have proved in principle an excellent tool for $T_i$ measurements, the results obtained were disheartening.

Three-electrode ion traps of honeycomb type were installed on the Cosmos 2 satellite launched in April 1962 for measuring $T_i$ (Afonin *et al.*, 1965; Gringauz *et al.*, 1965). Instead of external grids the traps were equipped with a group of parallel tubes, whose lengths were great compared to their cross-



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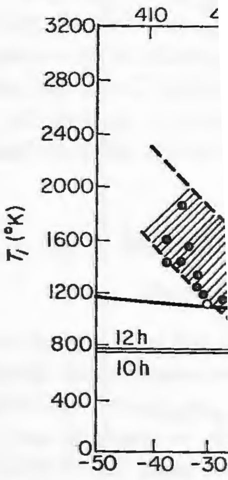


FIG. X.5. Measurements of T
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sections. The principle of measuring $T_i$ by means of such traps is similar to that of measuring $T_g$ by means of narrow tubes mentioned at the end of Section X.2. Since Cosmos 2 performed a complicated motion the determination of $T_i$ by the honeycomb traps was feasible in only a few cases when the orientation of the satellite happened to be suitable. The measured $T_i$ values relate to daytime conditions and to heights below 400 km. They are lower, by a factor of 2, approximately, than $T_e$ values measured under similar conditions by means of Langmuir probes. For instance,

when $H = 260$ km, $\quad T_i = 1300° \pm 200°$,

and

when $H = 300$ km, $\quad T_i = 1500° \pm 200°$.

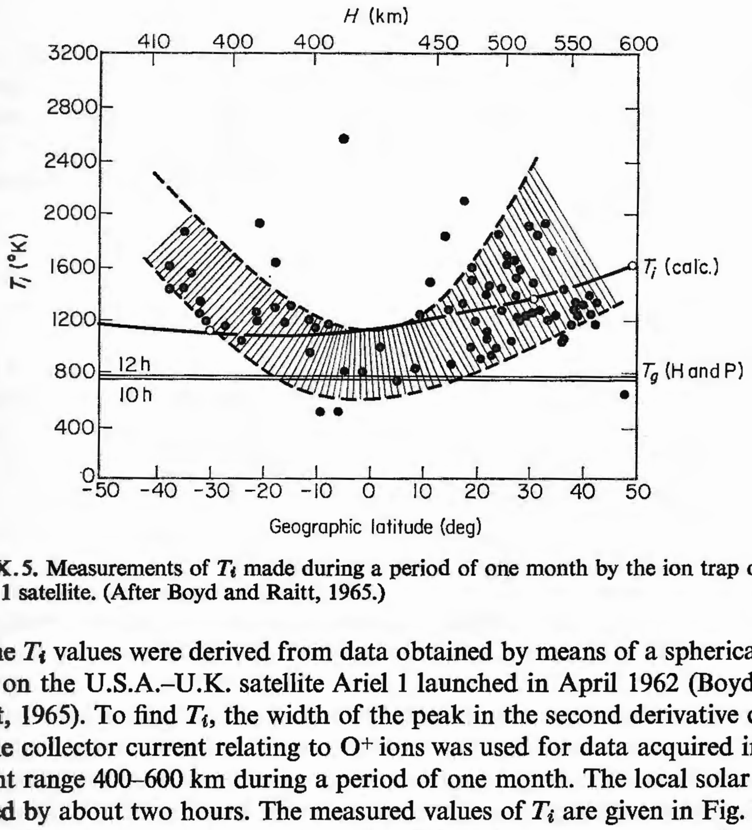


FIG. X.5. Measurements of $T_i$ made during a period of one month by the ion trap on the Ariel 1 satellite. (After Boyd and Raitt, 1965.)

The $T_i$ values were derived from data obtained by means of a spherical ion trap on the U.S.A.–U.K. satellite Ariel 1 launched in April 1962 (Boyd and Raitt, 1965). To find $T_i$, the width of the peak in the second derivative curve of the collector current relating to $O^+$ ions was used for data acquired in the height range 400–600 km during a period of one month. The local solar time varied by about two hours. The measured values of $T_i$ are given in Fig. X.5. The authors estimate that the error of each individual measurement does not exceed 200°, and the scatter of the $T_i$ values therefore appears to represent actual differences of temperature. These may be accounted for, partly by the fact that measurements were carried out at different local solar times, partly by seasonal changes in $T_i$, and by variations of magnetic activity. The main conclusions reached, however, are that for oxygen ions there are day-to-day variations of $T_i$ of a few hundred degrees, and that $T_i$ increases with increasing latitude. In Fig. X.5 the values of $T_g$ given by the Harris and Priester model for 10 and 12 h local time, respectively, are shown in the form of two straight lines parallel to the abscissa. Also shown in the same figure is a curve indicating the $T_i$ values computed by Willmore from the $T_e$ data obtained by Ariel 1 on the assumption that the neutral gas is heated by electrons as a

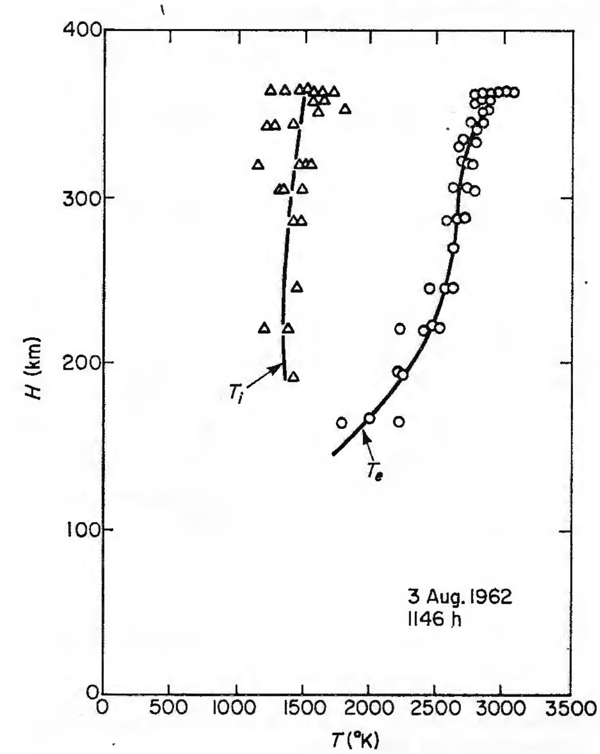


FIG. X.6. $T_i(H)$ and $T_e(H)$ profiles determined near noon on August 3rd, 1962. (After Nagy *et al.*, 1963.)

result of Coulomb interactions of electrons with ions. Boyd and Raitt (1965) have pointed out that, since the rate of energy exchange between particles essentially depends on the ion mass and concentration, the scatter of $T_i$ values observed in the experiment may be attributed to variations in ion density and ion composition.

Figure X.6 shows a $T_i(H)$ profile which was obtained by Nagy *et al.* (1963), simultaneously with a $T_e(H)$ profile, for heights between 180 and 365 km during daytime on 3 August 1962; the apparatus was separated from the rocket and consisted of a spherical ion trap and a Langmuir probe. The

$T_i(H)$ profile shows that, over the height range mentioned, $T_i$ was constant ($T_i \sim 1\,400°$) whereas $T_e$ increased with height and exceeded $T_i$ at all heights. It should be noted that the shape of $T_e(H)$ in Fig. X.6 is very different from the theoretical models proposed by Hanson (1963) and Dalgarno *et al.* (1963); the $T_e$ maximum near 220 km is absent.

It is worth noting that, judging by published data, no further attempts appear to have been made since 1962 to measure $T_i$ by direct methods for heights below 1 000 km.

Probably, the first estimate of the value of $T_i$ in the outermost ionosphere (for values of $H \leqslant 20\,000$ km) was made by Gringauz *et al.* (1960a) who stated that $T_i$ does not exceed tens of thousands of degrees.

Gringauz *et al.* (1966b) have suggested that, in order to obtain an estimate of $T_i$, use could be made of the collector current modulation produced in an ion trap, with its outer grid at zero potential, resulting from the rotation of the space vehicle. The application of this method of analysis to data obtained on the Electron 2 satellite (Gringauz *et. al.*, 1966b) has enabled the upper limit of possible $T_i$ values at heights from about 5 000 to 7 500 km to be reduced to 9 000–10 000°.

## X.4. Measurements of Electron Temperature, $T_e$

### X.4.1. *Vertical Distribution of* $T_e$ (*Rocket Measurements*)

It is certain that the vertical distribution of $T_e$ can be most reliably determined from measurements by rockets launched into nearly vertical trajectories. Among the first published results of such measurements of $T_e$ are those of Aono *et al.* (1961 and 1962), Brace *et al.* (1963) and Spencer *et al.* (1965). All these results show that, for heights up to about 150 km, the values of $T_e$ measured are usually low (about 1 000–1 200°K) and apparently close to $T_g$; occasionally, however, at these heights large $T_e$ values (up to about 2 000°K) are observed, indicating the evident absence of thermal equilibrium during these measurements. As the height increases, $T_e$ increases and reaches values close to 3 000°K by day (Brace *et al.*, 1963; Spencer *et al.*, 1965).

The absence of thermal equilibrium in the F region of the ionosphere was convincingly demonstrated by Spencer *et al.* (1965) in experiments with the thermospheric probes discussed in Section X.2. Figure X.7 gives the results of two measurements performed in 1963 by means of two rockets launched from Wallops Island, U.S.A. The rockets were launched at similar local times ($16^h$–$17^h$ approximately) on 18 April and on 20 July, 1963. The data shown in Fig. X.7 were obtained during the descent of the rockets. A comparison of the $T_e$ and $T_g$ values measured on 18 April clearly demonstrates the absence of thermal equilibrium in the F region of the ionosphere.

At $H$=250 km the $T_e$ value of about 2 000° is more than double the $T_g$ value.

The second pair of curves in Fig. X.7 shows the values obtained on 20 July 1963, and thus represents ionospheric conditions during a solar eclipse; during the rocket flight 85 to 75% of the sun's photosphere was eclipsed. These measurements, together with those made by rockets launched from Fort Churchill during the same eclipse (Smith *et al.*, 1965) provide excellent proof of the fact that solar ultraviolet radiation is the main source of heating the electrons in the ionospheric F region.

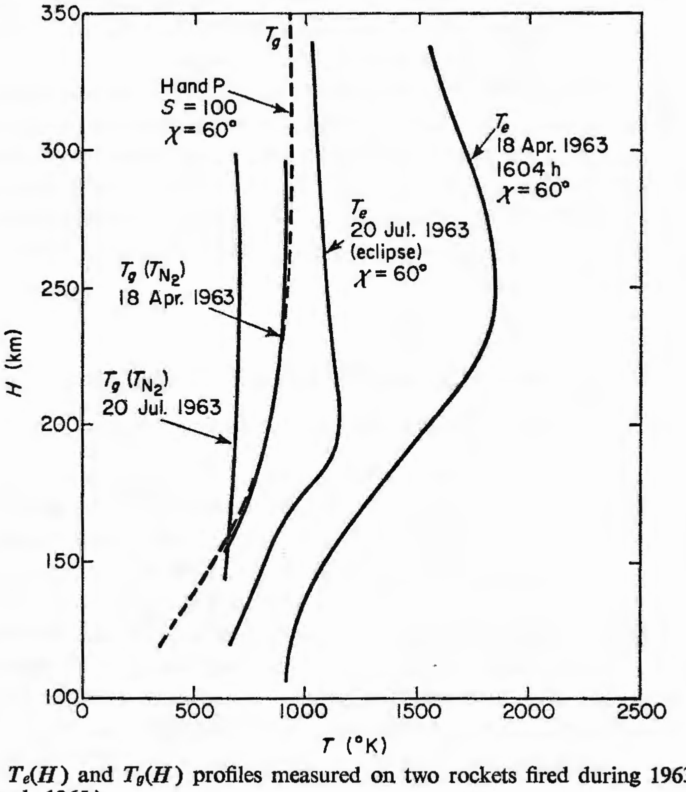


FIG. X.7. $T_e(H)$ and $T_g(H)$ profiles measured on two rockets fired during 1963. (After Spencer *et al.*, 1965.)

If the $T_e$ and $T_g$ values shown in Fig. X.7 for 18 April are compared with those of 20 July, it will be seen that the $T_e$ value during the eclipse was only about half that measured under normal conditions. It should be noted that the difference between the $T_g$ values was much smaller, the eclipse value of $T_g$ being slightly reduced. Below 150 km the difference between the $T_e$ values obtained on the two days is insignificant ($T_e$~1 000°K on both dates).

It can be shown that neither the divergence of the $T_e$ curves at heights above 150 km, nor their similarity lower down, are the result of the seasonal change, April to July. This is proved by the fact that Smith *et al.* (1965) observed the same effect in $T_e$ data obtained by Langmuir probes on four rockets, three of which were launched in an interval of about one hour from Fort Churchill during the same eclipse of 20 July, 1963. These data are shown in Fig. X.8; the maximum of the eclipse occurred at 2106 U.T., and the curve at the extreme right relates to a rocket launched after the end of the eclipse. It appears that, below about 150 km, the value of $T_e$ was very little affected by the eclipse (the error of measurement being $\pm 100°K$), while between 150 and 190 km the eclipse effect is more pronounced.

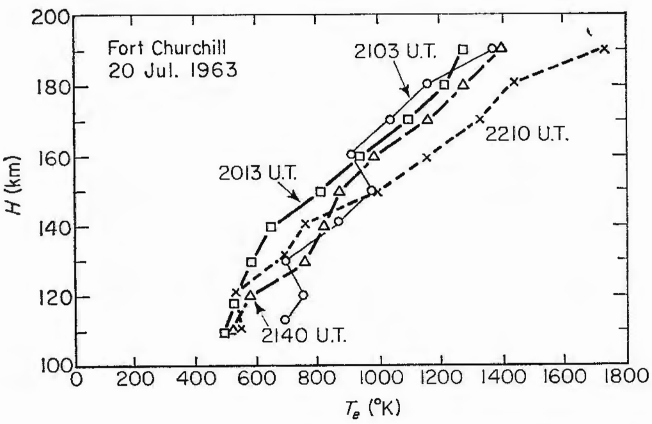


FIG. X.8. $T_e(H)$ profiles obtained from four rockets fired from Fort Churchill during the eclipse day 20 July 1963. (After Smith *et al.*, 1965.)

The measurements of Spencer *et al.* (1965) and Smith *et al.* (1965) prove conclusively that, as stated above, solar ultraviolet radiation is the main source of electron heating in the ionospheric F region, but not necessarily in the E region where other sources may also be involved, e.g. electric fields produced by the dynamo effect.

At the same time, the dependence of $N_e$ in the E region on the flux of the extreme ultraviolet radiation from the sun has long been known from ionospheric data, such as diurnal variations of critical frequencies or ionospheric observations during solar eclipses (e.g. Papaleksy, 1938). This dependence was clearly demonstrated (Smith *et al.*, 1965) during the above-mentioned rocket experiment, and the authors conclude that the daytime E region is mainly created by solar X rays. Thus solar radiation, being the main source of ionization, probably is not the main source of electron heating in the E region.

Of great interest are the results of $T_e$ measurements made on a rocket launched directly into the region of a visible aurora on 8 February, 1964 from Fort Churchill (Ulwick *et al.*, 1965). At heights of 300 to 320 km $T_e$

values of about 5 000° were observed; these values considerably exceed those usually found in the ionosphere.

Various $T_e(H)$ profiles obtained at middle latitudes by the Spencer and Brace group between 1961 and 1964 are shown in Fig. X.9; also included in the figure is the $T_e(H)$ curve from Fig. X.6 (Nagy *et al.*, 1963). Another daytime result which confirms the Hanson and Dalgarno theoretical models, was obtained during the firing of the L-3-1 rocket in Japan, at 1100 L.M.T. on 11 July, 1964, up to a height of 850 km; a maximum of $T_e$ was observed at

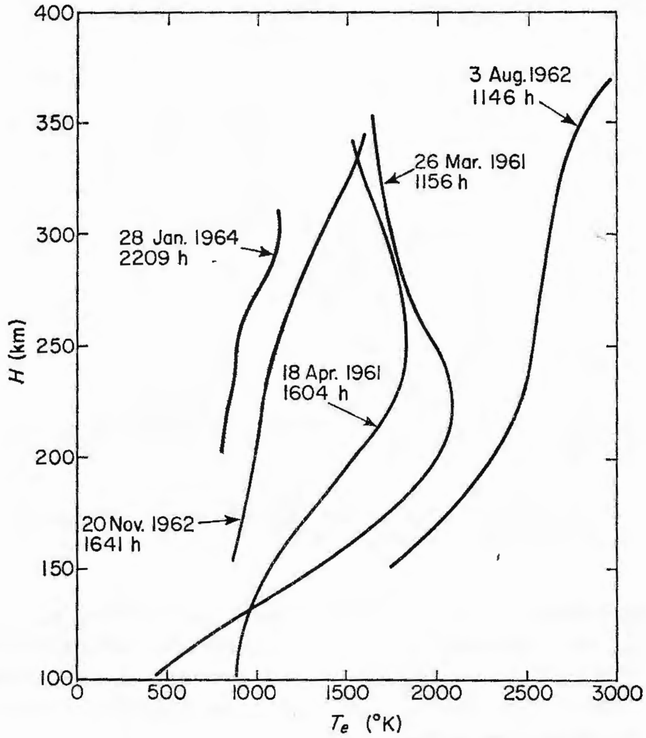


FIG. X.9. $T_e(H)$ profiles obtained by the Spencer and Brace group during the years 1961 to 1964. The curve for 3 August 1962, represents data reported by Nagy *et al.* (1963).

300 km [COSPAR Information Bulletin No. 27, p. 115 (1965)]. Other daytime results, however, have been obtained which contradict these models, and there are also $T_e(H)$ profiles for other periods of the day which have not yet been explained by theoretical calculations.

Authier *et al.* (1965) have reported a comparison of measurements of $T_g$, using the Na and AlO resonance glow technique, and of $T_e$, measured by Bourdeau by means of Langmuir probes; the measurements were made on rockets launched at twilight in the Sahara. It will be seen from Fig. X.10 that a pronounced minimum of $T_e$ was detected at a height of 275 km.

The $T_e(H)$ profile shown in Fig. X.11 was obtained by means of Langmuir probes on a U.S.S.R. geophysical rocket launched during a morning in September 1965 at a middle latitude (Gringauz *et al.*, 1966a). The data correspond to the ascent of the rocket, and each point represents the average of $T_e$ over a height interval of about 50 km. The solid line indicates the definite increase of $T_e$ with height, although the data points oscillate about the line. These results have been discussed in detail by Gdalevich *et al.* (1966).

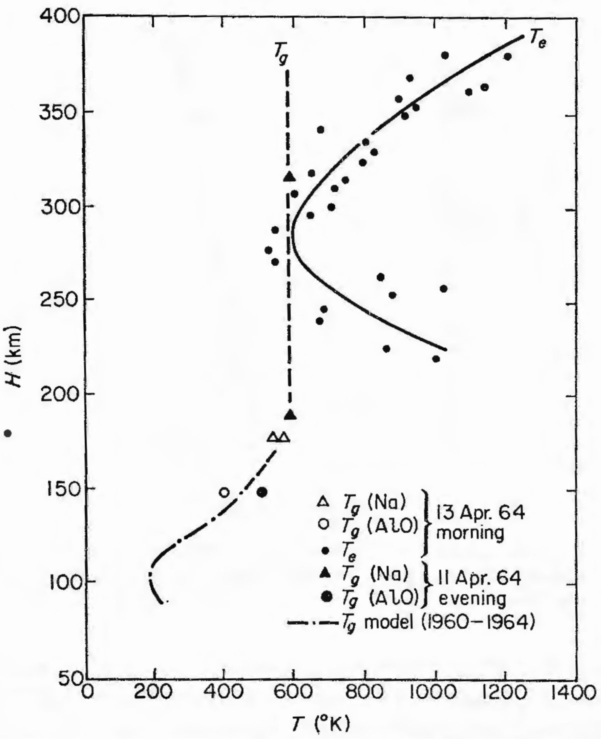


FIG. X.10. Measurements of $T_g$ made by Authier *et al.* (1965), and of $T_e$ made by Bourdeau on rockets launched at twilight in the Sahara during April 1964.

The results obtained in 1962 by means of Ariel 1 are discussed in the next section of this review, and they indicate an increase of $T_e$ with height at altitudes above 400 km (Bowen *et al.*, 1964b).

As already noted in Section X.1, Geisler and Bowhill (1965a) have shown that, if sufficient allowance is made for the thermal conductivity of the electron gas, a greater variety of $T_e(H)$ profiles can be explained than with previous models (Dalgarno *et al.*, 1963; Hanson, 1963). Geisler and Bowhill pointed out that the absence of a maximum of $T_e$ at heights of about 220 km is attributable to inefficient cooling of the electron gas which occurs near solar-minimum conditions. The considerable decrease in electron concentration during this period brings about a lowering of the efficiency of electron cooling at the expense of heat transfer to other particles.

Hirao (1966) has reported interesting $T_e$ measurements made by a rocket launched in Japan in August 1965 at about 11h local time, which reached a height of over 700 km. Measurements made by means of a high-frequency

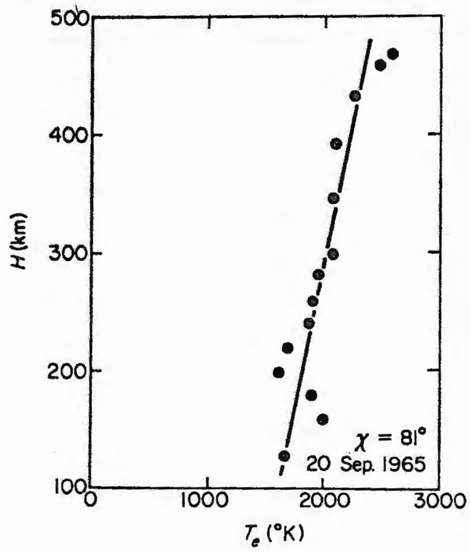


FIG. X.11. $T_e(H)$ data reported by Gringauz *et al.* (1966a) and by Gdalevich *et al.* (1966); the data were obtained by means of a rocket launched during a morning in September 1965 at middle latitudes.

probe showed that, besides the general tendency of $T_e$ to increase with height, there are present a number of maxima and minima in the $T_e(H)$ profile, that is a "stratification" of $T_e$, with the thickness of the "layer" (as defined by the distance between adjoining maxima and minima) of the order of 100 to 150 km (see Fig. X.12). A similar $T_e(H)$ profile was also obtained by another Japanese rocket launched in 1965 to a height of more than 300 km (Hirao, 1966), and Hirao has suggested that such variations of $T_e$ with height are caused not by the fluctuations of the heat source, but rather by variable heat losses apparently related to changes with height in the concentration and chemical composition of the ions and neutral particles. No aeronomic arguments substantiating the existence of such changes have, however, been advanced by Hirao. It is noteworthy, perhaps, that the oscillatory character of the data shown in Fig. X.11 resemble results shown in Fig. X.12.

We see, in conclusion, that many of the results of $T_e$ measurements made by means of rockets fail to agree with theoretical models of $T_e(H)$ developed up to the present time, and further experimental and theoretical investigations of the problem are certainly needed. Although rocket measurements have been performed in several parts of the world, such as Japan (Aono *et al.*, 1962; Hirao, 1966), the U.S.A. (Brace *et al.*, 1963; Nagy *et al.*, 1963; Spencer *et al.*, 1965), Canada (Smith *et al.*, 1965; Ulwick *et al.*, 1965), Algeria (Authier *et al.*, 1965) and the U.S.S.R. (Gringauz *et al.*, 1966a), the fact that

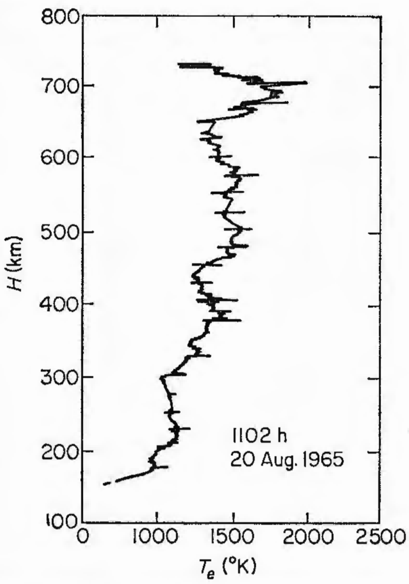


FIG. X.12. A $T_e(H)$ profile reported by Hirao (1966) which shows a number of maxima and minima of $T_e$ at different altitudes.

these measurements are insufficient in number and were not performed simultaneously prevents any conclusions being reached about the variation of $T_e$ with latitude and local time, etc. These variations may, however, be established on the basis of satellite measurements to be discussed in the next section.

### X.4.2. *The Variation of* $T_e$ *with Time and Latitude* (*Satellite Measurements*)

Table X.1 gives some information about the orbits of some satellites used for measurements of $T_e$ by probe methods.

In the $T_e$ probe measurements on Explorer 8 (Bourdeau, 1961; Bourdeau and Donley, 1964) the eccentricity of the orbit and the fact that measurements were performed only during direct radio contact with the earth forced the experimenters to assume that $T_e$ was independent of height. This has simplified the analysis of the data obtained, and the most important and reliable result of the experiments has been the detection of a pronounced peak in $T_e$ (up to $2{\cdot}5 \times T_g$) near sunrise.

TABLE X.1. SATELLITES USED FOR $T_e$ MEASUREMENTS

| Name | Launch Date | Perigee (km) | Apogee (km) | Inclination to Equator (degrees) |
|---|---|---|---|---|
| Explorer 8 | 3 December 1960 | 425 | 2400 | 50 |
| Cosmos 2 | 7 April 1962 | 212 | 1540 | 49 |
| Ariel 1 | 26 April 1962 | 400 | 1200 | 54 |
| — | June 1962 | 260 | 317 | 75 |
| — | July 1962 | 160 | 181 | 75 |
| Explorer 17 | 3 April 1963 | 258 | 920 | 58 |
| Explorer 22 | 9 October 1964 | 1000 | 1000 | 80 |
| IMP 2 | 4 October 1964 | 200 | 95000 | 34 |

With Cosmos 2, $T_e$ measurements were performed at heights between 212 and 550 km in daytime during the periods of direct radio contact with the satellite (Afonin *et al.*, 1965; Gringauz *et al.*, 1965). Unfortunately, insufficient data were available to permit the authors to separate the effects of latitudinal changes of $T_e$ from those produced by other parameters. The daytime $T_e$ values measured in the F region of the ionosphere were found to lie between 1 800 and 3 000°. At those points where $T_i$ was measured simultaneously, $T_e$ exceeded $T_i$ by a factor of between 2 and 2·5, thus indicating the absence of thermal equilibrium.

$T_e$ measurements from Ariel 1 were carried out by means of two planar Langmuir probes, with the results being stored for the whole orbit of the satellite (Bowen *et al.*, 1964b; Willmore, 1965). Although a considerable amount of data could be obtained in this way, simultaneous changes of height, latitude and local solar time along the orbit made it difficult to assess separately the individual effects of any of these factors on $T_e$. In an attempt to determine these individual effects, the original data were subjected to a complicated statistical analysis based on the assumption that seasonal changes in $T_e$ were negligible over a four-month period. $T_e$ data relating to individual revolutions of the satellite were not published, but the results of the primary data analysis for Ariel 1 measurements from 28 April to 22 August, 1962, show that $T_e$ increases with height over the whole height range investigated at all local times. This result does not agree with the Hanson and Dalgarno theoretical models, according to which no increase of $T_e$ with height should occur at the altitudes concerned in the daytime ionosphere. The Geisler and Bowhill model, however, can account for the observed effect.

The published Ariel 1 data also indicate that $T_e$ increases with geomagnetic latitude. Bowen *et al.* (1964) state that an increase in solar ultraviolet radiation will produce an increase in $T_e$. In all instances an inverse correlation between $T_e$ and electron concentration $N_e$ has been found, which can be fully explained by the theories of Hanson and Dalgarno and is attributable to changes in the conditions of electron cooling due to a decrease of $N_e$.

It should be noted that the maximum $T_e$ values recorded in the published Ariel 1 data only slightly exceed 2 000°K, whereas measurements from rockets (see Section X.4.1.) and from other satellites frequently yield $T_e$ values of about 3 000°. This point will be discussed below.

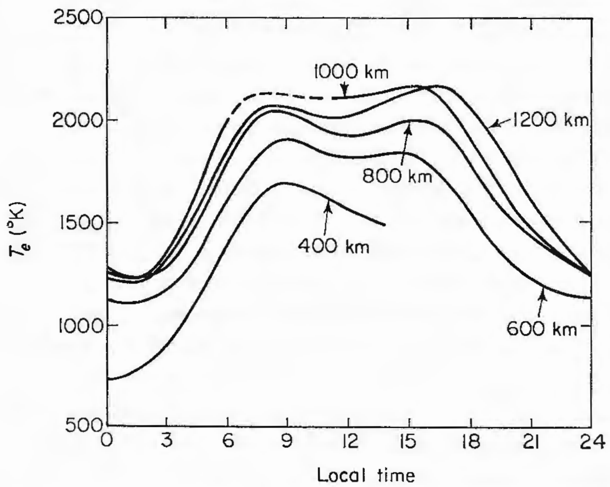


FIG. X.13. $T_e$ values measured at different altitudes and local times by means of the Ariel 1 satellite. (After Willmore, 1965.)

According to Bowen *et al.* (1964b) the diurnal variation of $T_e$ at different heights did not show the peak near sunrise detected by Explorer 8. Subsequently, however, the data were re-examined by Willmore (1965) who observed a sunrise peak and showed that its value decreased with height (see Fig. X.13).

Bowen *et al.*, and Willmore have drawn attention to the considerable dependence of $T_e$, at heights between 400 and 1 200 km, on geomagnetic latitude. An increase in $N_e$ during magnetic storms is always accompanied by a decrease in $T_e$. The variations of $T_e$ during magnetic storms take place along magnetic shells. It should be noted that, according to the early rocket measurements performed by Spencer *et al.* (1962) and Brace *et al.* (1963), magnetic disturbances do not cause a decrease, but an increase in $T_e$.

Near a geomagnetic latitude of about 50° a weakly developed maximum of $T_e$ has been detected.

Willmore (1965) also suggests that the negative correlation between $T_e$ and $N_e$ at night, as well as the change in the $T_e$ height gradient which occurs at about 600 km (near the height where the ion composition alters) show that the temperature variations are associated with changes in the rate of electron cooling through collisions, and indicate the existence of a heating mechanism under night conditions. This is particularly pronounced at latitudes above about 30°.

At about the time when the $T_e$ measurements were being made by Ariel 1 a series of short-term measurements were performed at lower altitudes by means of two satellites whose approximate orbital parameters are listed in Table X.1 (Sagalyn *et al.*, 1965). No exact dates are given by the authors who only state that the data relate to the period June–July 1962. Spherical probes screened by grids were used in these measurements which also revealed a distinct diurnal variation of $T_e$ with a peak near sunrise. The daytime value of $T_e$ at heights from 250 to 300 km was about 3 000°.

Further measurements have been performed by means of cylindrical Langmuir probes on Explorer 17 and published by Brace *et al.* (1965); these mainly cover the period 4 April–10 July, 1963, geographical latitudes from 30° and 50° and the height range between 260 and 550 km. In addition, some data are given relating to geomagnetic latitudes of 10° and 60°. Because of the eccentricity of the orbit of Explorer 17 Brace *et al.* have made some simplifying assumptions in their analysis of the results in order to determine the dependence of $T_e$ on various factors.

The main conclusions of Brace *et al.* about the variation of $T_e$ with height and with geomagnetic latitude, about the negative correlation of $T_e$ with $N_e$, and about the existence of a night-time source of ionospheric heating, all agree qualitatively with those of the Ariel 1 experimenters. There are, however, some quantitative discrepancies; for instance, the morning peak of $T_e$ is 2 700° and considerably exceeds the magnitude determined by the Ariel 1 measurements.

Brace *et al.* (1965) agree with Willmore about the existence of an energy source to produce the night-time difference between $T_e$ and $T_g$, and calculate that, in order to explain the increased values of $T_e$ measured by Explorer 17 at a height of 400 km, a heat input of about 20 eV $cm^{-3}$ $sec^{-1}$ is required. This is five times the magnitude of the input needed to explain the Ariel 1 data. Brace *et al.* also believe that a flux of electrons with energies of about 100 eV, corresponding to an energy flux of $20^{-2}$ ergs $cm^{-2}$ $sec^{-1}$ with a heating efficiency of 10% may explain the observed values of $T_e$ without conflicting with the data of other geophysical observations.

Explorer 22, launched in 1964, has a circular orbit with a large inclination (about 80°). The nearly total absence of height variations, and the rapid changes of latitude combined with slow changes in longitude make it an almost ideal vehicle for investigating the latitude variations of ionospheric parameters. The first results of measurements carried out on this satellite by means of Langmuir probes similar to those used on Explorer 17 (Brace *et al.*, 1965) and in rocket experiments with thermospheric probes (Spencer *et al.*, 1965) have been published by Brace and Reddy (1965). Two recent papers (Brace and Miller, 1966; Brace *et al.*, 1966) discuss in detail the results obtained at high and low latitudes. It should be noted that, contrary to earlier data from satellites which reached lower latitudes, the Explorer 22 data indicated the existence of distinct latitudinal maxima of $T_e$ instead of a monotonic increase of $T_e$ with latitude. Figure X.14 shows some of the Explorer 22 data.

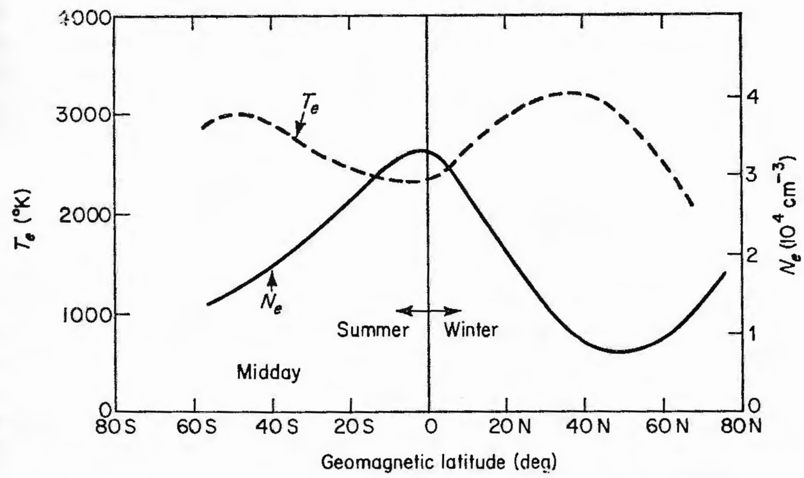


FIG. X.14. Latitude variations of $T_e$ and $N_e$ measured by means of the Explorer 22 satellite. (After Brace and Reddy, 1965.)

Figure X.15 shows the diurnal variations of $T_e$ and $N_e$ derived from Explorer 22 data (Brace and Reddy, 1965) and of $T_e$ from Ariel 1 data (Willmore, 1965); the data correspond to a height of about 1 000 km and a geomagnetic latitude of 40°. It will be noticed that the daytime values of $T_e$ based on Explorer 22 measurements exceed the corresponding values from Ariel 1 by approximately 1 000°.

It is difficult to explain the low values of $T_e$ measured by Ariel 1, since the $S$ values were higher in 1962 than in 1964, and according to Willmore (1965) $T_e$ should increase with $S$. The discrepancy may be explained by a decrease in $N_e$ which appears to occur at all heights in the ionosphere as the solar activity decreases (shown, for example, by a comparison of $N_i$ data obtained by Sputnik 3 and Cosmos 2 (Gringauz *et al.*, 1964), so that the resulting deterioration in the conditions for electron cooling causes $T_e$ to increase. Another possible explanation of the discrepancy may arise in the processing of the Ariel 1 data, particularly in the averaging processes; this is difficult to assess, since "individual" data relating to the separate revolutions of Ariel 1 have not been published.

On all the satellites mentioned above, various types of modified probes were used for the detection of charged particles. Some information about $T_e$ and $T_i$, however, can be obtained by processing ionograms produced by ionosondes aboard topside sounder satellites of the Alouette type.

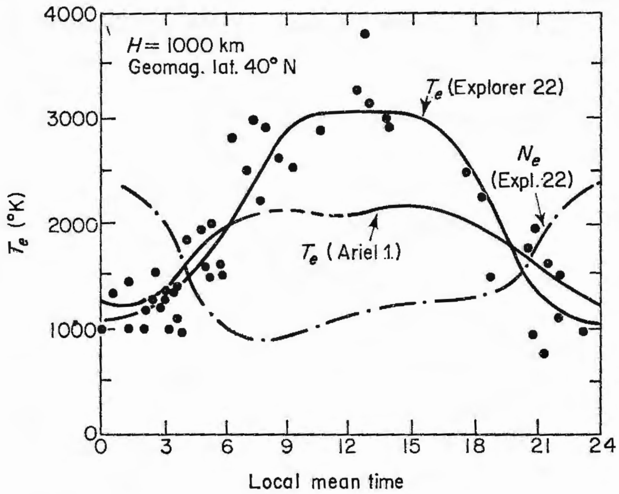


FIG. X.15. Diurnal variations of $T_e$ observed by Explorer 22 and Ariel 1, and also of $N_e$ observed by Explorer 22. All the data relate to a height of 1 000 km and 40° geomagnetic latitude. (After Brace and Reddy, 1965; Willmore, 1965.)

Topside sounder data show that $N_e$ decreases monotonically with height above the F-region maximum. If the value of $N_e$ is given by

$$N_e = N_0 \exp\left[-H/\bar{H}_e\right],$$

where $N_0$ is the electron density at zero reference height, then we may call $\bar{H}_e$ the "plasma scale height".

Watt (1965) has demonstrated that, if the atmosphere is in diffusive equilibrium

$$\bar{H}_e = \frac{T_e + T_i}{\partial T_e/\partial H + \partial T_i/\partial H + \bar{m}_i\, g/k},$$

where $\bar{m}_i$ is the mean ion mass, $g$ the acceleration of gravity, and $k$ Boltzmann's constant.

Only $\bar{H}_e$ is obtained in the experiment and it will prove difficult to determine the parameters $T_e$, $T_i$, $\partial T_e/\partial H$, $\partial T_i/\partial H$ and $\bar{m}_i$.

In deriving $\bar{H}_e$ from Alouette 1 ionograms for the height range 400–900 km and geomagnetic latitudes from 48° N to 78° N, Watt made simplifying assumptions about (a) the absence of horizontal $N_e$ gradients, (b) the existence of diffusive equilibrium over the whole height range considered, and (c) the thermal equilibrium of ions among themselves. On the basis of Hanson's theoretical considerations, $T_i$ was assumed to equal $T_e$ at a height of 800 km, and account was taken of the ion mass data obtained from rocket measurements by Taylor *et al.* (1963). Watt calculated $T_e$ and $T_i$ as a function of latitude in the height range 500–800 km for daytime and night-time conditions in summer 1963 and winter 1963/64. Watt (1965) has processed a number of Alouette 1 ionograms and, although the assumptions made are reasonable, it should be recognized that the accuracy and reliability of the $T_e$ and $T_i$ values obtained in this way are considerably lower than that of the $N_e$ values determined from these ionograms, or of $T_e$ and $T_i$ measurements by probes. With regard to the latitudinal distribution of $T_e$ and $T_i$, Watt (1965) has reported maxima at high latitudes, which resembled those maxima observed in $T_e$ in Explorer 22 data.

### X.4.3. $T_e$ *in the Outermost Ionosphere*

The only experimental measurements of $T_e$ in the outermost ionosphere (much higher than 1 000 km) were performed with the IMP-2 satellite by means of a retarding-potential method using a three-electrode planar trap (Serbu and Maier, 1966). On the basis of data obtained during half-a-year the authors conclude that $T_e$ increases with height in proportion to $R^2$ (where $R$ is the geocentric distance) and $N_e$ decreases in proportion to $R^{-3}$ up to $R=5\,R_E$ where $R_E$ is the earth radius. From about $5\,R_E$ to the apogee at 15·9 $R_E$, $T_e$ remains almost constant at 1 to 2 eV, and $N_e$, too, only changes slightly.

The results of Serbu and Maier are very interesting and impressive; so far only a small part of the available data appears to have been published and, although Serbu and Maier (1966) have left some questions unanswered, it is expected that more will be explained in later publications. For example, the accuracy of determining $T_e$ by probe techniques is rather uncertain when $T_e$ is only 0·1 to 0·2 eV while the retarding potential varies in discrete steps differing by about 1V. It is also not clear why none of the radial $N_e$ distributions obtained by Serbu and Maier show a "knee", that is a sudden decrease of $N_e$ (or $N_i$), near 4·5 $R_E$ as has been repeatedly found by three other independent groups of observers (Carpenter, 1966; Bezrukikh and Gringauz, 1965; Taylor *et al.*, 1965) using different methods.

However, in spite of these two questions (which may be answered by Serbu and Maier in further publications) the fact that there appears to be a considerable increase in $T_e$ with height in the outermost ionosphere is very interesting.

No theoretical calculations of $T_e(H)$ at heights of the order of several $R_E$

have been carried out. The only theoretical paper dealing with this problem is one by Geisler and Bowhill (1965b), in which consideration was given to the change, along a geomagnetic tube, in the temperature of the ionospheric plasma heated by fast photoelectrons ascending up the tube from the F region. Calculations for a geomagnetic tube crossing the level of 1 000 km at about 40° geomagnetic latitude have given an almost isothermal $T_e$ distribution along the total length of the tube; the temperatures are estimated to be about 3 000°K at solar minimum. These calculations do not contradict the findings of Serbu and Maier, but neither do they predict the observed increase in $T_e$ with height.

Energetic photoelectrons can, of course, ascend along geomagnetic lines of force with little energy loss since the cross-section of their interaction with ions decreases approximately as $v_e^{-4}$ (where $v_e$ is the electron velocity). Perhaps they contribute to the high plasma temperatures in the outlying regions of the ionosphere. The possibility cannot be ruled out that there is some thermal coupling between this outer ionosphere and solar wind plasma which penetrates into the magnetosphere by some mechanism such as that suggested by Dessler and Walters (1964) or Block (1966).

Clearly, new experimental and theoretical investigations will be necessary before an understanding is reached of the sources of heat in the outermost ionosphere.

## X.5. Conclusions

The present paper has not dealt with any of the valuable temperature data obtained by means of ground-based observations, for example those reported by Evans (1965a, c); the ground-based observations have been considered fully in Chapter IX. It is, in any event, interesting to find what conclusions can be reached about the temperatures of ionospheric particles solely on the basis of rocket and satellite measurements.

Direct measurements of $T_g$ obtained by means of mass-spectrometers, and observations of the fluorescence of K, Na and AlO are the most reliable, but they are not particularly numerous. Above about 100 or 200 km all the measurements show that $T_g$ increases with height. The dependence of $T_g$ on the intensity of solar ultraviolet radiation (as characterized by the 10·7 cm solar radiation flux $S$) apparently becomes more pronounced with increasing height (up to about 300 km). $T_g$ values measured by direct methods are found to agree quite well with those from theoretical models based on observations of satellite drag.

Some of the $T_g$ measurements discussed either confirm the existence of an isothermal zone at about 200–300 km, e.g. those based on mass-spectrometer experiments (Pokhunkov, 1965; Spencer *et al.*, 1965), while other measurements do not disprove the existence of the region, e.g. measurements of resonance-line broadening (Authier *et al.*, 1965; Blamont and Chanin-Lory, 1965). The results indicate how the height and temperature of this zone depend on $S$ and on local time. The data are not, however, sufficient to establish that the zone has a permanent existence. Evidence has been provided (Mikhnevich, 1965) about deviations from isothermal conditions over the above-mentioned height range, even though the deviations may be short-lived and therefore not detectable in results averaged over long periods. $T_g$ values at the height of the isothermal zone, as determined by Explorer 17, are characterized by considerable scatter, the reasons for which are not quite clear.

Rocket and satellite measurements of $T_i$ are very scarce. The daytime values of $T_i$ are much lower than the $T_e$ values at heights from 200 to 400 km (Nagy *et al.*, 1963; Afonin *et al.*, 1965), and also at heights from 400 to 600 km (Boyd and Raitt, 1965) where the scatter of $T_i$ values amounts to many hundreds of degrees from day to day. $T_i$ tends to increase with latitude (Boyd and Raitt, 1965). At heights from about 5 000 to 8 000 km $T_i$ is lower than 9 000–10 000° (Gringauz *et al.*, 1966b). No direct measurements of $T_i$ and $T_g$ have been performed simultaneously.

Rocket and satellite measurements of $T_e$ are much more numerous than those of $T_g$ and $T_i$. At heights above about 200 km thermal equilibrium is absent in the daytime ionosphere ($T_e > T_i > T_g$) (Brace *et al.*, 1963; Bourdeau and Donley, 1964; Bowen *et al.*, 1964a; Gringauz *et al.*, 1965, and others). In the night-time ionosphere thermal equilibrium is also absent (Brace *et al.*, 1965; Willmore, 1965). Daytime $T_e(H)$ profiles were obtained in 1961, with maxima at heights of about 220 km (Brace *et al.*, 1963) in close agreement with theoretical models (Dalgarno *et al.*, 1963; Hanson, 1963). During 1962 and subsequently, most day-time measurement at a height of about 220 km show an increase of $T_e$ with height (Nagy *et al.*, 1963; Gringauz *et al.*, 1966a; Hirao, 1966) although in 1963 a $T_e(H)$ profile was obtained with a maximum at about 250 km (Spencer *et al.*, 1965). It may be that the shape of the $T_e(H)$ profile at heights below about 1 000 km varies with the solar cycle, as theory would appear to indicate (Geisler and Bowhill, 1965a); measurements of $T_e(H)$ during the next solar maximum should thus be very interesting.

Rocket measurements of $T_e(H)$ during the solar eclipse of 20th July, 1963, and simultaneous rocket measurements of solar ultraviolet radiation at different parts of the spectrum, have proved convincingly that solar radiation is the heat source for electrons in the daytime F region. On the other hand, the electrons in the E region produced by solar X rays appear to be subjected to other sources of heating which may include, for example, electric fields (Smith *et al.*, 1965; Spencer *et al.*, 1965).

The existence, near sunrise, of a peak in the diurnal variation has been established by Bourdeau and Donley (1964), Sagalyn *et al.* (1965) and Willmore (1965); this peak decreases as the height increases (Willmore, 1965).

In most measurements an inverse correlation of $T_e$ with $N_e$ has been observed (Brace and Reddy, 1965; Willmore, 1965).

According to rocket measurements (Brace *et al.*, 1963) magnetic disturbances produce an increase in $T_e$, whereas the Ariel 1 data indicate a decrease of $T_e$ (Willmore, 1965). No definite conclusions about the influence of magnetic disturbances on $T_e$ can be reached on the basis of the published rocket and satellite results.

The variation of $T_e$ with geomagnetic latitude by day at a height of about 1 000 km shows a minimum near the equator and maxima near geomagnetic latitudes of about 40°N and 50°S. At night-time these maxima are displaced towards higher geomagnetic latitudes (Brace and Reddy, 1965).

It has been shown that $T_e$ in the outermost ionosphere (at $H \gg 1\,000$ km) increases in proportion to $R^2$ (where $R$ is the geocentric distance) reaching values of 10 000–20 000° at heights of about 25 000 km (Serbu and Maier, 1966). Theoretical models of $T_e(H)$ for these heights have not yet been published.

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