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PLANETARY AERONOMY XIII: ELECTRON AND ION TEMPERATURES IN THE IONOSPHERE

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ABSTRACT

The relationships between the temperatures of the electrons, positive ions and neutral particles of the ionosphere are discussed. It is argued that on the basis of the heating due to solar ultraviolet radiation, the electron and ion temperatures will become equal at high altitudes and that both tend to increase above the neutral particle temperature. Anomalously high values of the electron temperature, accompanied by a red glow, are predicted to occur in the region of 400 km during the dawn period. The expected diurnal and latitudinal variations of the electron temperature are described.

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ELECTRON AND ION TEMPERATURES IN THE IONOSPHERE

A. Dalgarno

1. Introduction

The relationships between the temperatures of the electrons, the positive ions and the neutral particles which constitute the ionosphere are fundamental to the interpretation of its behavior. The theoretical studies of the effect of solar ultraviolet radiation (Hanson and Johnson 1961; Hanson 1963; Dalgarno, McElroy and Moffett 1963) show that departures from temperature equality are to be expected at altitudes above about 150 km, the difference between the electron temperature and the heavy particle temperature attaining a maximum somewhat greater than 1000^{o}K at an altitude of about 220 km and decreasing rapidly as the altitude increases to 300 km. These conclusions are in substantial agreement with the Langmuir probe measurements (cf. Bourdeau 1963). It has been suggested (Dalgarno *et al*. 1963) that greater differences may occur near sunrise.

According to the theoretical studies, the difference between the electron temperature T_e and the positive ion temperature T_i should decrease rather slowly with increasing altitude above the peak of the F region. The difference may persist over several hundred kilometers, the predicted magnitude of T_e-T_i depending sensitively upon the assumed density of the ambient electrons. The behavior of T_i with increasing altitude has not been investigated except for the hypothetical situation in which T_e is taken as constant at 1800°K and the neutral particle temperature T_n is taken as constant at 1400°K. According to Hanson (1963), T_i remains equal to T_n up to an altitude of 600 km and then increases to T_e as the altitude increases to 1000 km.

Measurements of T_e and T_i at altitudes above 300 km are in apparent contradiction. Bourdeau (1963) has shown that rocket and satellite observations at mid-latitudes in quiet conditions can be satisfactorily explained assuming that T_e, T_i and T_n are approximately equal, a conclusion that is in harmony with the observations of backscatter reported by Bowles, Ochs and Green (1962), but not with the observations of backscatter by Evans (1962) who deduces that $T_e \sim 1.6\ T_i$ between 300 and 700 km or of Pineo and Hynek (1962) who deduce that $T_e \sim 2\ T_i$.

2. Electron Temperatures in the Upper F Region

Provided that $T_e < 3\ T_i$, the electrons located above the peak of the F region cool by collisions with the positive ions. If O^+ is the major positive ion, the equilibrium electron temperature T_e is determined by the equation

$$\frac{Q}{n_e^2} = 5.5 \times 10^{-7}\ (T_e - T_i) \Big/ T_e^{3/2} \quad . \tag{1}$$

where Q is the heat flux density in eV cm^{-3} sec^{-1} and n_e is the electron number density. According to the calculations of Dalgarno, McElroy and Moffett (1963), the values of Q/n_e^2 which occur in the upper F region, if it is assumed that the heating takes place where the solar ultraviolet radiation is absorbed, are of the order of 10^{-9}.

The solutions of (1) for heat flux densities of the order of $10^{-9}\ n_e^2$ are shown in Figure 1 for ion temperatures of 1000°K, 1500°K and 2000°K.

It is clear from Fig. 1 that T_e-T_i is very sensitive to the values adopted for T_i, Q and n_e^2 when Q/n_e^2 is of the order of 10^{-9}. As a specific example, Table 1 lists the electron-ion temperature differences T_e-T_i corresponding to various possible electron densities for a heat flux density of 500 eV cm^{-3} sec^{-1} which is appropriate to an altitude of about 400 km during midday.

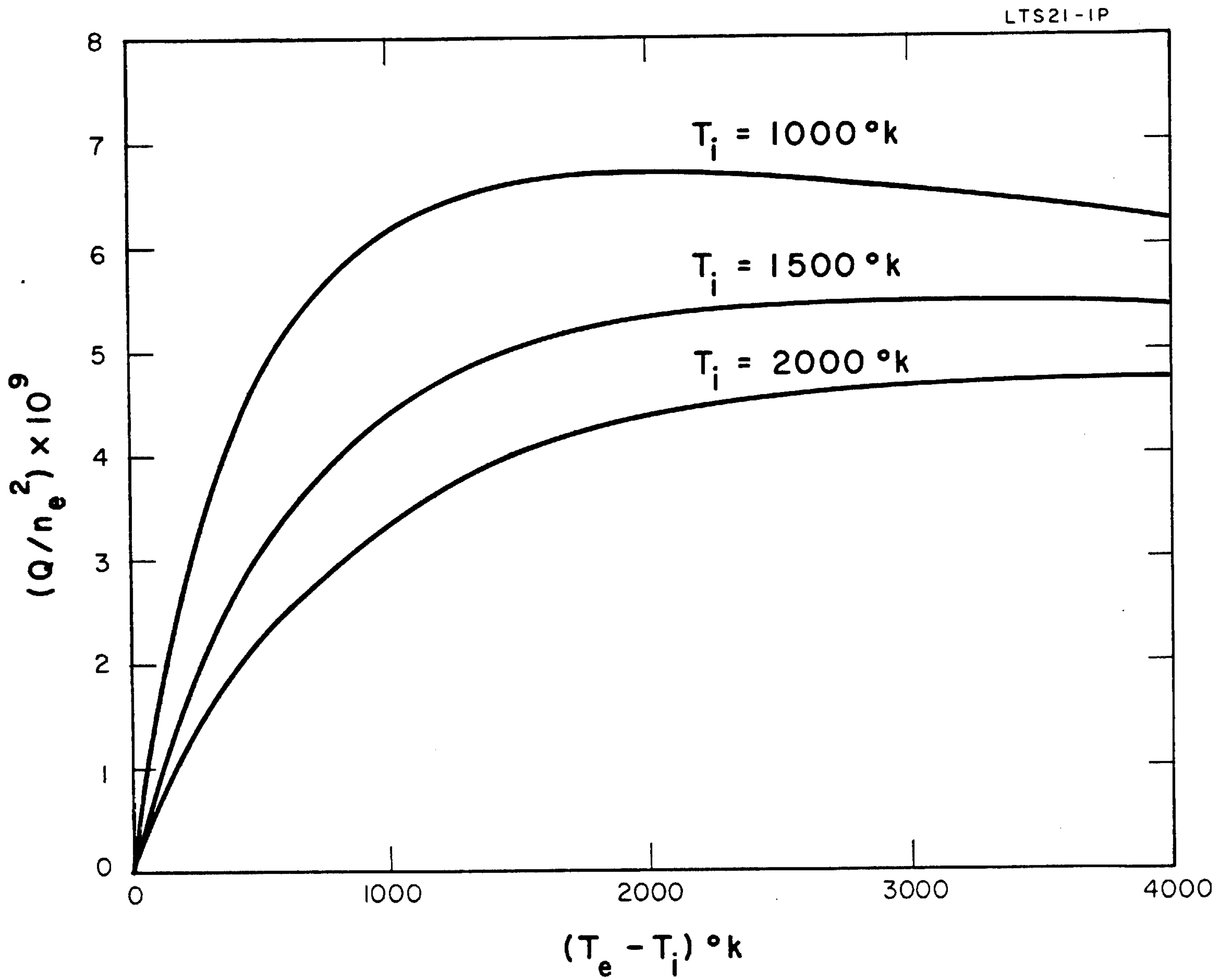

Figure 1. Equilibrium electron temperatures for various heat flux densities.

Table 1

Temperature Differences ($T_e - T_i$) at 400 km for Various Assumed Ambient Electron Densities and Ion Temperatures Corresponding to a Heat Flux Density of 500 eV cm^{-3} sec^{-1}

$n_e \times 10^{-6}$ cm^{-3} / T_i	$1000^{o}K$	$1500^{o}K$	$2000^{o}K$
1.4	10	30	50
1.2	20	40	65
1.0	30	60	95
0.8	50	90	140
0.6	90	175	280
0.4	220	510	880
0.35	380	860	1640
0.325	490	1200	4000
0.300	720	3000	--
0.275	2000	--	--

With the assumption of local heating, the heat flux density in the upper F region where atomic oxygen is the major constituent is approximately

$$Q = 3 \times 10^{-6}\ n(O)\ \mathrm{eV\ cm^{-3}\ sec^{-1}} \qquad (2)$$

where n(O) is the atomic oxygen number density. The altitude distribution of atomic oxygen may be represented by the expression of diffusive equilibrium

$$n\ (O/z) = n\ (O/r)\ \exp\ (-\ z/H) \qquad (3)$$

where H is the scale height and z is the altitude measured with respect to some reference level r at which n(O) = n (O/r). If the ions and electrons are also in diffusive equilibrium, then similarly

$$n_e(z) = n_e(r)\ \exp\ (-\ z/H') \qquad (4)$$

where the ion and electron scale height H' in an atmosphere of O, O^+ and e is related to H by

$$H' = \frac{T_e + T_i}{T_n} H \quad . \qquad (5)$$

It follows that in the upper F region,

$$\frac{Q}{n_e^2} = \frac{3 \times 10^{-6}\ n(O/r)}{n_e(r)^2} \exp \left\{ -\ \frac{z}{H} \left(\frac{T_e + T_i - 2T_n}{T_e + T_i} \right) \right\} \quad . \qquad (6)$$

If $T_n = T_i$, (6) simplifies to

$$\frac{Q}{n_e^{\;2}} = \frac{3\times10^{-6}\, n(O/r)}{n_e(r)^2} \exp\left\{ - \frac{z}{H}\left(\frac{T_e - T_i}{T_e + T_i}\right)\right\} , \tag{7}$$

which implies that with increasing altitude, T_e and T_i will ultimately become equal. However, as Hanson (1963) notes, a small difference between T_e and T_i can persist over a considerable height range. Thus, if we assume that T_e/T_i is 1.5 at an altitude of 400 km, (7) yields the ratios listed in Table 2.

Table 2

Ratios of T_e/T_i According to (7)

Altitude (km)	400	400+H	400+2H	400+3H	400+4H
T_e/T_i	1.50	1.37	1.32	1.28	1.24

Since H is at least 75 km, (7) shows that T_e/T_i should decrease to 1.2 at an altitude of 700 km. The interpretations of the backscatter observations of Evans (1962) and of Pineo and Hynek (1962) in terms of a constant value of T_e/T_i between 300 and 700 km can not be understood on the basis of direct solar ultraviolet heating and an ionosphere in diffusive equilibrium. It is necessary to postulate either an additional heating mechanism or an electron distribution which decreases more rapidly with increasing altitude. This conclusion is strengthened by the fact that helium and hydrogen ultimately become the major constituents with the consequence that the efficiency of heating is decreased and the efficiency of cooling is increased.

The temperate difference $T_e - T_i$ may actually decrease more rapidly than Table 2 suggests since there is a possibility that T_i is larger than T_n at great altitudes so that (6) should be used in place of (7).

3. Ion Temperatures in the Upper F Region

At high altitudes, the positive O^+ ions are heated by collisions with the electrons and cooled by collisions with the oxygen atoms, and

$$\frac{dT_i}{dt} = \frac{5.5\times10^{-7} n_e(T_e - T_i)}{T_e^{3/2}} - 8.6\times10^{-14} n(O) (T_i - T_n), \quad (8)$$

since the collision frequency of O^+ in O is about 10^{-9} n(O) $\sec^{-1}$. Therefore,

$$\frac{d}{dt} (T_e + T_i) = \frac{Q}{n_e} - 8.6\times10^{-14} n(O) (T_i - T_n) \quad . \quad (9)$$

In equilibrium,

$$T_i - T_n = \frac{1.16\times10^{13}}{n(O)} \frac{Q}{n_e} \quad . \quad (10)$$

Now using (4),

$$T_i - T_n = \frac{3.5\times10^{7}}{n_e} \quad . \quad (11)$$

This represents an upper limit since the collision frequency of O^+ in O is actually a slowly increasing function of temperature (Dalgarno 1961). It suggests that at 500 km, the ion temperature may exceed the neutral particle temperature by 100°K and at 800 km by 1000°K.

The coefficient in (11) is computed on the assumption that the heating is produced locally.† Since the major part of the heating due

†Hanson (1963) has shown that the fast photoelectrons produced at great altitude may escape from the ionosphere.

to solar ultraviolet radiation is provided by collisions of fast photoelectrons with the ambient electrons, the assumption of local heating is unlikely to be correct. Nevertheless, there is some local heating due to the photoionization processes which directly populate the ^2D metastable state of O^+. The O^+ (^2D) has a long radiative lifetime and above 200 km, it is probably deactivated by superelastic collisions

$$O^+(^2D) + e \rightarrow O^+(^4S) + e \quad . \tag{12}$$

According to Dalgarno and McElroy (1963), (12) gives rise to a heat flux density of approximately

$$Q = 6 \times 10^{-7} \; n(O) \text{ eV cm}^{-3} \text{ sec}^{-1} \; , \tag{13}$$

which in the absence of heat conduction and diffusion cooling is sufficient to maintain T_i some 200°K above T_n at 800 km and some 500°K at 1000 km.

The tendency of T_i to rise above T_n increases the difficulty presented by some of the backscatter data and emphasizes the discrepancy between them and the temperatures derived from charged particle density profiles. Thus, if the heating were such that $T_e = 1.6\ T_i$ at high altitudes, then it follows from (1) and (10) that

$$T_i - T_n = \frac{2\text{x}10^6 \; n_e}{T_i^{\frac{1}{2}} \; n(O)} \quad . \tag{14}$$

Adopting a model atmosphere employed by Hanson (1963), it follows that at 500 km, $T_n = 1200^{o}K$, $T_i = 1500^{o}K$ and $T_e = 2400^{o}K$; and at 700 km, $T_n = 1200^{o}K$, $T_i = 2100^{o}K$ and $T_e = 3400^{o}K$.

4. Diurnal Variation of Electron Temperatures

Provided Q/n_e^2 does not exceed 10^{-9} eV cm^{-3} sec^{-1}, $T_e = T_i$ and the diurnal variation of T_e must be very similar to that of T_n. If Q/n_e^2 does exceed 10^{-9} eV cm^{-3} sec^{-1}, it is more difficult to predict the diurnal variation of T_e since it depends sensitively upon the diurnal variations of the electron and neutral particle densities. According to (7), in the upper F region $T_e - T_i$ should decrease more slowly with increasing altitude as T_n increases, but the absolute value of $T_e - T_i$ at any given altitude depends also upon $n(O/r) / n_e(r)^2$, the variation of which with increasing T_n is uncertain. In any event, the electron number density probably responds more slowly to an increase in T_n than does the neutral particle density and high values of T_e/T_i may occur during periods of geomagnetic activity. This, rather than any additional preferential heating of the electrons, may be the explanation of the observations of Spencer et al. (1962).

The diurnal variation of T_e in the lower F region is of special interest since anomalously large values may occur near sunrise. It has been suggested, in particular, that T_e may exceed $3T_i$ in which case the electron temperature increases to a new equilibrium value determined by the efficiency of cooling by collisions with the neutral particles (Dalgarno et al. 1963).

Preliminary calculations indicate that the heat flux density at dawn at an altitude of 400 km is about 700 eV cm^{-3} sec^{-1}. The resulting values of $T_e - T_i$ are shown in Table 3 for a range of concentrations n_e.

Table 3

Dawn Values of T_e-T_i

n_e (cm^{-3})	$1x10^6$	$8x10^5$	$6x10^5$	$4x10^5$	$3.2x10^5$
$(T_e-T_i)^oK$	50	75	130	1430	2000

If the ambient electron density at 400 km is less than $3.2x10^5$ cm^{-3}, and it probably is, T_e increases above $3T_i$ and may attain a value of several thousand degrees. The principal limiting factor may be the increase in electron density.

The cooling at the high equilibrium temperatures is effected partly by collisions leading to excitation of the ^1D state of atomic oxygen, and the runaway of the electrons should become evident through the appearance of a red glow.

During the dawn period, there should occur also discontinuities in T_e as a function of altitude.* Explicit calculations are in progress to predict the altitude range of the anomalously high values of T_e.

Similar effects should occur at greater altitudes during dawn twilight. Thus, the diurnal variation of T_e-T_i expected from solar

*The examination of this possibility was stimulated by a conversation with Dr. P. Molmud and Dr. S. Altshuler.

ultraviolet heating at midlatitudes consists of very high values in the region of 400 km at dawn, the magnitude and altitude of the maximum temperature decreasing with increasing time.

During the afternoon, the magnitude and altitude should increase but because the ambient electron concentration is larger, the very high dawn values are not expected to occur at sunset. After sunset, kinetic equilibrium should be rapidly established.

The diurnal variation will be less marked at high latitudes where diffusion is not inhibited by the magnetic field, especially since the faster electrons diffuse more rapidly than the slower electrons.

There is some question as to whether the dawn behavior at mid-latitudes is a stable configuration. It may be that the sunrise effect is manifested by a sharp reduction in electron density rather than by a sharp increase in electron temperature.

A departure from temperature equilibrium at dawn is suggested by a preliminary analysis of Ariel data (Willmore, Boyd and Bowen 1962) and by the backscatter ofservations of Bowles et al. (1962), but not by the backscatter observations of Evans (1962). There is no evidence that T_e exceeds $3T_i$ at any altitude for a short period about dawn, but this may be due to the averaging procedures which have been applied to the data.

5. Latitude Variation of Electron Temperatures

Spencer, Brace and Carignan (1962) find from rocketborne Langmuir probes that T_e is higher in auroral latitudes than at midlatitudes in quiet conditions, and the main characteristics of the Ariel data (Willmore et al. 1962, Bourdeau 1963) are a general rise in T_e and a less pronounced diurnal variation at high latitudes. It may be that the enhanced temperature is caused by corpuscular radiation (which would give rise to some preferential heating of the electrons), but higher values of T_e can be explained qualitatively as a consequence of a latitudinal decrease in the ambient electron density. The measurements of Spencer et al. (1962) show that the charged particle densities at auroral latitudes can be much lower than those occurring at midlatitudes in quiet conditions.

Willmore et al. (1962) have remarked that the Ariel data are consistent with significant atmospheric heating by particles dumped at high latitudes, but the satellite drag data do not show any important variation with latitude except during magnetically disturbed conditions (Jacchia 1963). As Bourdeau (1963) has shown, there are several uncertainties in the preliminary analysis of the Ariel electron temperatures.

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