

HYDROGEN IONS*

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Whistler observations which first indicated surprisingly large electron densities at several earth radii (cf. "Whistlers and Exospheric Structure" by L. R. O. Storey) led to the interpretation that the outer ionosphere consists predominantly of hydrogen ions (protons) and electrons. Consequently, the name "protonosphere" has been suggested for this region of the earth's outer atmosphere (Johnson, 1960). The hydrogen ions of the protonosphere are now considered to be the result of charge exchange between ionospheric oxygen ions and neutral hydrogen of the telluric geocorona (Johnson, 1961), rather than due to photoionization. While the protonosphere was originally envisioned (Johnson, 1960) to be a direct extension of the ionospheric F region, where oxygen ions predominate, it has been established recently that a region of helium ions whose extent varies with temperature and solar cycle, intervenes between the regular ionosphere and the protonosphere (cf. "Helium Ions" by W. B. Hanson).

It was first pointed out by Dungey (1955), that the charge exchange process



may be of importance in the upper atmosphere. Johnson, (1960) was the first to argue that this process represents the major source of hydrogen ions constituting the protonosphere. Since oxygen and hydrogen have nearly the same ionization potential, reaction (1), which is of the accidental asymmetric resonance type, occurs about as rapidly either way and therefore presents the source and sink for H^+ . It has been suggested (Rapp, 1963) that the cross-section for process (1)

is large enough (10^{-15} to 10^{-16}cm^2) so that (1) occurs rapidly enough to be the most important chemical reaction involving hydrogen ions in the upper atmosphere. (In interplanetary space, photoionization and radiative recombination obviously will become the controlling processes.) Hanson, Patterson and Degaonkar (1963) have recently inferred from experimental data a rate coefficient $K = 3.9 \times 10^{-10} \text{cm}^3 \text{sec}^{-1}$ for reaction (1), corresponding to a reaction cross-section of $7.6 \times 10^{-16} \text{cm}^2$; these values are considered to be correct within a factor of three.

The altitude distribution of an ion, such as H^+ , is governed by the equation of continuity

$$\frac{\partial(H^+)}{\partial t} = q(H^+) - L(H^+) - \text{div } F(H^+) \quad (2)$$

where $q(H^+)$ is the production rate and $L(H^+)$ is the loss rate of hydrogen ions and $F(H^+)$ is the flux of H^+ resulting from diffusion under gravity.

For the steady state, $\frac{\partial(H^+)}{\partial t} = 0$, and considering mainly an upward or downward flux of hydrogen ions (i.e. in the z direction), equation (2) can be expressed by

$$q(H^+) - L(H^+) - \frac{\partial F(H^+)}{\partial z} = 0 \quad (3)$$

While equation (3) is not readily amenable to an analytic solution, but has to be solved numerically (cf. Bates and Patterson, 1961; Hanson, 1963, Hanson et al, 1963) there are two simple limiting cases to (3): 1) where diffusion of hydrogen ions can be neglected, *chemical equilibrium* ($q = L$) will prevail, and 2) at high enough altitudes where diffusion is the predominant process, *diffusive*

equilibrium will prevail ($\frac{\partial F(H^+)}{\partial z} = 0$)

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1. Chemical Equilibrium Distribution of H^+

Noting that both production and loss of H^+ are controlled by the charge exchange reaction (1), we can write (cf. Bates and Patterson, 1961)

$$Kn(H)n(O^+) = \sigma Kn(H^+)n(O) \quad (4)$$

where K is the rate coefficient and $\sigma = 8/9$ is the value for the ratio of the products of the statistical weights. Thus, the chemical equilibrium distribution for hydrogen ions is given by:

$$n(H^+) = \frac{9}{8} \frac{n(H)}{n(O)} n(O^+) \quad (5)$$

where $n(X)$ is the number density of the neutral constituent X which is distributed exponentially according to its scale height $\underline{H}(X) = kT/m(X)g$, with k Boltzmann's constant, T the absolute temperature, $m(X)$ the mass of the constituent and g the acceleration of gravity; the number density of oxygen ions, $n(O^+)$ is distributed according to its scale height (assuming predominance of O^+) $\underline{H}(O^+) = k(T_e + T_i)/m(O)g$, where T_e and T_i are the electron and ion temperature, respectively, which may be different from each other. The distribution above a reference level h_0 where $n(H^+) = n_0(H^+) = 9n_0(H)n_0(O^+)/8n_0(O)$ and above which all constituents are distributed exponentially, is given by

$$n(H^+) = n_0(H^+) \exp \{z/\underline{H}(7)\} \quad (6)$$

where Z is the geopotential altitude above the reference level ($z = \int_{h_0}^h (g/g_0) dh$) which takes into account the variation of the acceleration of gravity g with altitude, and $\underline{H}(7)$ is the scale height corresponding to an effective mass of 7 AMU, (resulting from the combined scaled heights), and to the acceleration of gravity at the earth's surface g_0 . (It is assumed that all constituents entering into the R.H.S. of (5) are in diffusive and thermal equilibrium; e.g., $\underline{H}(O^+) = 2\underline{H}(O)$). Thus, it is obvious that H^+ in chemical equilibrium is *increasing* with altitude.

2. Diffusive Equilibrium Distribution of H^+

It is now generally known that the diffusive equilibrium distribution of an ion (such as H^+) in a mixture is not independent of the other ionic con-

stituents because of the presence of an electrostatic field which is set up to prevent further charge separation between electrons and ions diffusing under gravity. This electrostatic field, which depends on the mean ionic mass of the mixture and on the electron and ion temperatures, is sensed by all ions in the mixture and tends to counteract the gravitational force. This fact has been known to astrophysicists for many years (cf. Eddington, 1926), but has been revived more recently (Dungey, 1955; Mange, 1960) in connection with the terrestrial ionosphere. According to this concept, the distribution of H^+ in diffusive equilibrium is given by

$$n(H^+) = n_0(H^+) \exp \left[- \int_0^z \left(\frac{1}{\underline{H}(H^+)} - \frac{\tau}{1+\tau} \frac{m_+ g_0}{kT_i} \right) dz \right] \quad (7)$$

where $\underline{H}(H^+) = \frac{kT_i}{m(H^+)g_0}$, $\tau = T_e/T_i$ is the ratio of electron to ion temperature and $m_+ = \sum m_i n_i / \sum n_i$ is the mean ionic mass of the mixture. Since the upper ionosphere consists of a ternary ion mixture (H^+ , He^+ , O^+) we also can write (7) as

$$n(H^+) = n_0(H^+) \exp \left[- \int_0^z \frac{dz}{\underline{H}(H^+)} \right] \quad (8)$$

where $\underline{H}(H^+) = \frac{\underline{H}(H)(1+\tau)}{n(O^+) + n(He^+) + n(H^+)}$
 $\frac{n(O^+) + n(He^+) + n(H^+)}{n(H^+) - \tau[14n(O^+) + 2n(He^+)]}$

Thus, it is easily seen that, e.g. for the case of thermal equilibrium ($\tau = 1$), $\underline{H}(H^+) < O$, if $n(H^+) < 2[n(He^+) + 7n(O^+)]$ and approaches $2\underline{H}(H)$ when H^+ becomes the predominant ion. As long as H^+ is a minor constituent, it is *increasing* with altitude. In fact, in a predominantly O^+ ionosphere the initial increase of H^+ in diffusive equilibrium follows almost the same scale height as the one corresponding to chemical equilibrium (6) and in this case, the predominance of chemical or diffusive equilibrium is undistinguishable from the behavior of the altitude distribution alone. Figure 1 shows the behavior of the distribution of H^+ in diffusive (and thermal) equilibrium in the presence of He^+ and O^+ at a temperature of $T = 1200^\circ K$. For the simple computation of the diffusive and thermal equilibrium distribution of an ion in a mixture, the following formula is quite

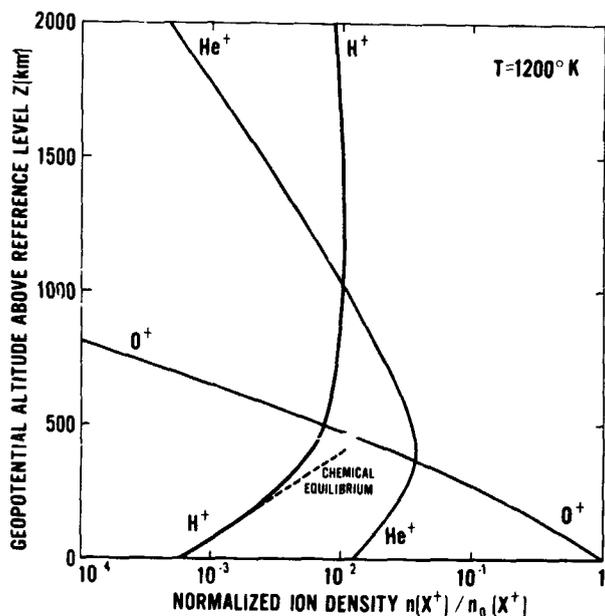


FIGURE 1

useful (cf. Hanson, 1962; Bauer, 1962):

$$n(X_k^+) = n_0(X_k^+) \exp \left[-z/H_k \right] \cdot \left[\frac{\sum n_0(X^+)}{\sum n_0(X^+) \exp(-z/H_i)} \right]^{1/2} \quad (9)$$

(Although the electron temperature often differs from the ion temperature, calculations based on thermal equilibrium generally can provide an adequate representation of the altitude distribution of an ion in diffusive equilibrium.)

The actual distribution of hydrogen ions in the upper atmosphere is one which more closely corresponds to a steady-state distribution (3) with a non-vanishing flux of ions. Figure 2 shows an observed H^+ distribution (Taylor, Brace, Brinton and Smith, 1963) which has been interpreted by Hanson et al, (1963) in terms of a steady-state distribution, corresponding to a rate coefficient for the charge exchange reaction of $3.9 \times 10^{-10} \text{cm}^3 \text{sec}^{-1}$ and an upward flux of hydrogen ions at 900 km of $F(H^+) = 1.3 \times 10^7 \text{cm}^{-2} \text{sec}^{-1}$.

The altitude where diffusion becomes predominant compared to chemical processes in controlling the distribution of H^+ can be estimated by comparing the rate of the downward flux of hydrogen ions $\left(-\frac{\partial \Gamma(H^+)}{\partial z} \right)$, with the loss rate due to charge

exchange with atomic oxygen (cf. Bates and Patterson, 1961) leading to the criterion

$$Kn(O) = \frac{D(H^+, O^+)}{2[H(O)]^2} \quad (10)$$

where $D(H^+, O^+) = 8.4 \times 10^{15} (T/1500^\circ)^{5/2} / [n(H^+) + n(O^+)]$, (Hanson, 1963) is the diffusion coefficient for Coulomb diffusion of protons through a predominantly O^+ ionosphere, and the other quantities have their previously defined meaning. For the condition representing the data of Taylor et al, (1963) shown in Figure 2, the transition from predominantly chemical to diffusive control would occur at an altitude of about 600 km and since the diffusion process is controlled by oxygen ions one would expect that one scale height, $H(O^+) = 2 H(O)$, above and below this level diffusive and chemical equilibrium respectively, would prevail. This estimate is in good agreement with the conclusions of an analysis of the behavior of the ratio of $n(H^+)/n(He^+)$ of the same data by Bauer (1964). It was first shown by Hanson and Ortenburger (1961) that oxygen ions represent an effective barrier to the diffusion of H^+ because of Coulomb interaction between ions. Thus, the "critical level" defined by the criterion

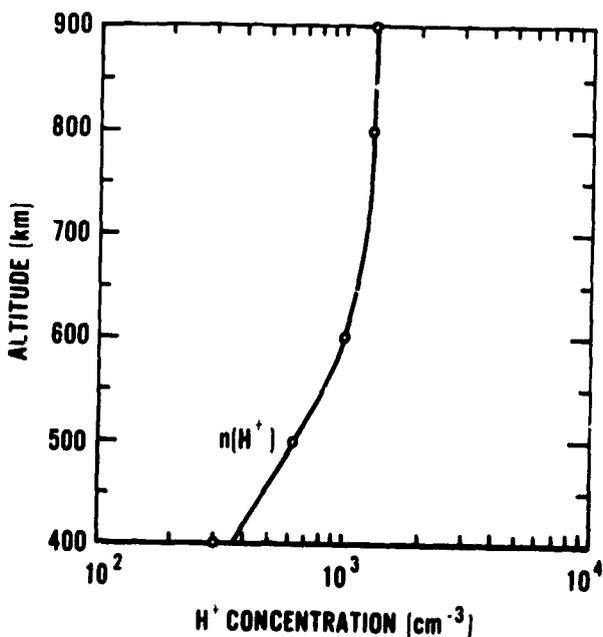


FIGURE 2

in (10) may be considered the base of the protonosphere.

Since the charge exchange process (1) controls the hydrogen ion concentration at the lower altitudes, the time-dependent boundary condition for the proton distribution at higher altitudes is given to a first approximation by the ratio of $n(H)/n(O)$ (cf. (5)). It is well known that the neutral hydrogen concentration is quite sensitive to the atmospheric temperature because of the thermal escape of neutral hydrogen. Thus, the concentration of hydrogen at a fixed level of say 500 km will be higher at low temperatures than at high temperatures. Consequently, the ratio $n(H)/n(O)$ at 500 km may vary by a factor of 100 for a temperature change of about $1000^\circ K$ (Bates and Patterson, 1961; Kockarts and Nicolet, 1963). Considering a ternary ion mixture (O^+ , He^+ and H^+) and boundary conditions varying with temperature according to the variation of the neutral species corresponding to these ions at 500 km (Bauer, 1963; Kockarts and Nicolet, 1963) the altitude domain where H^+ is the predominant ion can be described qualitatively as shown in Figure 3.

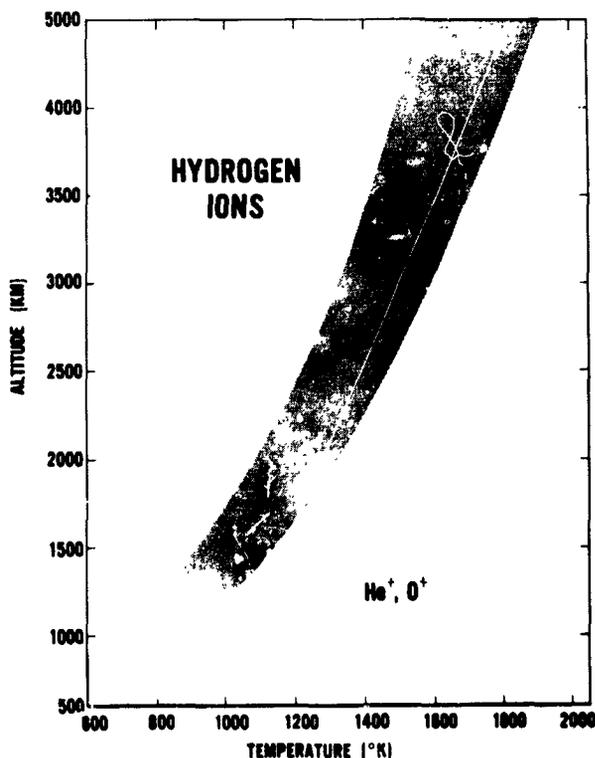


FIGURE 3

While such a concept is roughly indicative of the longer term temperature variations it may not be representative for diurnal variations, since the response time of the change in the hydrogen concentration to a change in temperature, at least at the lower temperatures, is of the order of tens of hours (Bates and Patterson, 1961). Although there is some indirect evidence of large diurnal variation in the concentration of hydrogen based on nighttime profiles of ion and electron density (Donley, 1963; King et al, 1963) which indicate a preponderance of H^+ already at altitudes below 1000 km, a large diurnal variation is difficult to explain in view of the long response time of hydrogen. Hanson and Ortenburger, (1961) have pointed out that because of the inhibiting effect of Coulomb diffusion, the protons cannot follow diurnal variations of the oxygen ions in the ionospheric F region and thus, the protonosphere is essentially decoupled from the upper ionosphere. In fact, the diurnal variation in the protonosphere based on whistler observations is only of the order of at most 30 percent. Since the total content of H^+ throughout the protonosphere is estimated to be about 7.5×10^{12} protons cm^{-2} (Hanson and Ortenburger, 1961), a 30 percent variation would imply an average flux of 5×10^7 ions $cm^{-2}sec^{-1}$ into the ionospheric F region during a 12-hour period. This flux would be, in general, upward during the day and downward during the night. The upward flux during the daytime, however, seems to be limited by the flux of neutral hydrogen from the lower atmosphere which is of the order of 2.5×10^7 atoms $cm^{-2}sec^{-1}$ (cf. Hanson and Patterson, 1963). As mentioned before, an upward flux of 1.3×10^7 protons $cm^{-2}sec^{-1}$ has been inferred by Hanson et al (1963) from the daytime proton distribution measured by Taylor et al (1963), (cf. Figure 2). Consequently, it appears (Hanson and Patterson, 1963) that, although nighttime conditions may allow a sufficient flux of protons to account for a 30 percent change in the hydrogen ion content of the protonosphere, the return flow required during the day does not appear to be feasible. Thus, it also seems unlikely that fluxes of protons between hemispheres affect the ion density at the $F2$ peak as has been suggested (Rothwell, 1962) since the observed proton fluxes are small compared with the normal flux through the $F2$ peak

which is estimated to be of the order of 2×10^8 to 10^9 ions $\text{cm}^{-2}\text{sec}^{-1}$ (Hanson and Patterson, 1963).

While the general behavior of hydrogen ions is already reasonably well understood, quantitative data especially those concerning its time variation are still inadequate. However, this situation will hopefully be remedied as the result of additional measurements with rockets and satellites.

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*Suggested as source for a more comprehensive treatment of the subject of this chapter.

MEAN DIURNAL VARIATION OF THE TOPSIDE IONOSPHERE AT MID-LATITUDES

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About 500 ionograms obtained with the Alouette satellite have been used to construct the mean diurnal variation of the mid-latitude topside ionosphere. Since this diurnal variation is based on observations from October to December 1962, quasisesonal effects had to be removed. The data are presented in form of electron density contours at fixed altitudes for two latitude ranges 35°N–40°N and 40°N–45°N. These two latitude ranges, which are approximately separated at the 75°W meridian by the 70° magnetic dip line, show remarkable differences in the detailed behavior of electron density as a function of time and altitude. From the scale heights of these mean distributions, it is apparent that there is absence of thermal equilibrium, at least during the day, and that light ionic constituents become important in the altitude region from 500 km to 1000 km, especially during the night. In addition, there is the suggestion of latitude gradients in the electron temperature and mean ionic mass.

INTRODUCTION

Prior to the launching of the Canadian topside sounder satellite *Alouette*, our knowledge of the vertical structure of the ionosphere above the F_2 peak was based on occasional rocket flights and incoherent radar backscatter observations from the ground, both of which are limited to specific locations. The topside sounder satellite provides us for the first time with a tool for the study of the topside ionosphere and its variation with latitude and time. Since the topside sounder operates only on command, observations are restricted to times when the satellite is within appropriate range of a telemetry and command station. We have made use of soundings of the topside ionosphere taken while the satellite was within range of the NASA telemetry station at Blossom Point, Maryland. The latitude coverage obtainable from this station is between 20°N and 55°N. We have concentrated on the study of ionograms obtained at mid-latitudes (35°N to 45°N), for which our data were most complete. As the result of the orbital characteristics of the Alouette satellite and combining north- and southward passes, about three months are required to obtain a complete diurnal variation of the ionospheric parameters. This paper presents the mean diurnal

behavior of the mid-latitude topside ionosphere for the three-month period from October to December 1962.

DIURNAL VARIATION OF ELECTRON DENSITY AT CONSTANT ALTITUDES

About 500 ionograms taken on magnetically quiet days during the first three months of Alouette's operation and covering the latitude range from 30°N to 50°N (and longitudes 50°W to 100°W) were selected for the present study. The ionograms were converted to electron-density profiles by means of the exponential lamination method described by *Fitzenreiter and Blumle*, (1964). For the study of the diurnal behavior of the topside ionosphere, a presentation in terms of electron density at constant altitudes was chosen. In this way a mean behavior could best be defined from the large number of data points. It became apparent that because of the latitude variation of the topside ionosphere, latitude ranges of not more than 5° latitude should be used, otherwise the spread in the electron density data points at constant altitudes would often be greater than the separation of electron density contours at successive 100 km altitude increments. Since below 35°N and above 45°N the data available were not sufficient to define a complete diurnal variation, only the two latitude ranges from 35°N to 40°N

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and from 40° to 45°N will be presented in this paper. The "mean" contours of electron density at 100 km height intervals are based on individual data points; the maximum spread of individual data points around the density contours at fixed heights is of the order of ± 25 percent. Figure 1 shows an example of such mean contours every 200 km up to the satellite altitude for the latitude range 40°N to 45°N. It can be seen that these mean contours have some oscillations superimposed which need not represent actual diurnal behavior, since the entire "diurnal" variation is based on three months of data. Thus, the mean

sity at the indicated local times. The smoothed variation of $S_{10.7}$ is plotted in the top portion of Figure 1 and it is evident that the electron density at fixed altitudes behaves in a similar fashion. At first, it may seem surprising that the topside electron density should be closely related to the 10.7 cm flux from the sun, since there is no obvious correlation of this parameter with the density at the F₂ peak, or below. However, it should be realized that at lower altitudes where production and loss of ionization play a pre-dominant role, the detailed correlation of electron density with short term solar variations as indicated by the 10.7 cm flux may be obscured by the fact that both, production and loss are affected by the variation of the appropriate neutral constituents and is thus effectively cancelled out. At altitudes above the F₂ peak, where the distribution is controlled by diffusion, and thus by the scale height, this effect may appear more directly since the scale height and temperature are known to be directly correlated with the 10.7 cm flux. Correspondingly, the "seasonal" effect superimposed on the diurnal variation was removed using the behavior of the

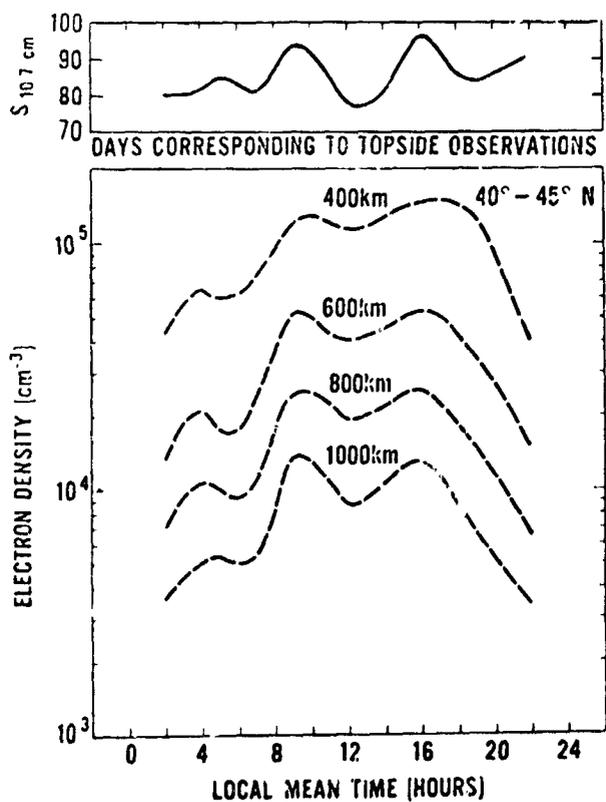


FIGURE 1.—Mean diurnal variation of electron density at constant altitudes based on actual data points and flux of 10.7 cm solar radiation (in units of $10^{-22} W/m^2/cps$) at times corresponding to the ionospheric measurements, indicating "quasi-seasonal" effects (27 day cycle).

contours presented in Figure 1 will also include quasi-seasonal effects, e.g., the 27 day cycle. This becomes quiet apparent, if the flux of the solar 10.7 cm radiation, $S_{10.7}$, which is an indicator of solar activity, is plotted for the days corresponding to the individual data points for electron den-

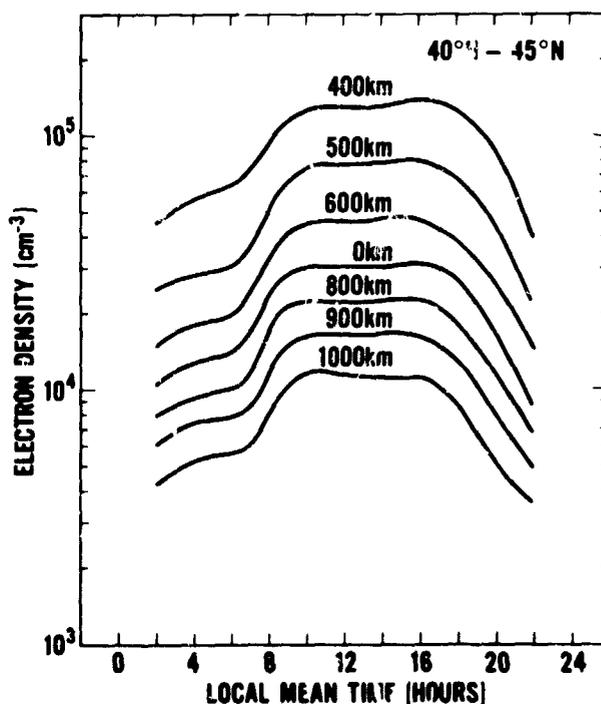


FIGURE 2.—Diurnal variation of electron density at constant altitudes for the latitude range 40° N to 45° N. (Corrected for quasi-seasonal effects).

10.7 cm solar flux as an indicator and normalizing the mean diurnal variation to a condition of $S_{10.7} = 85 \times 10^{-22} \text{ W/m}^2/\text{cps}$. This was accomplished by making a percentage correction of the electron density contours appropriate to the variation of $S_{10.7}$. The mean diurnal variation of electron density at fixed altitudes, corrected for quasi-seasonal effects, is shown in Figure 2 for the latitude range 35°N to 40°N and in Figure 3 for

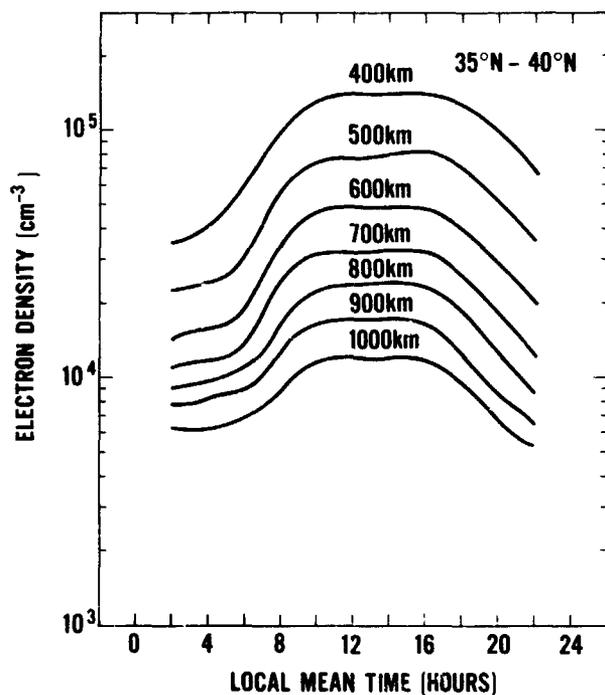


FIGURE 3.—Diurnal variation of electron density at constant altitudes for the latitude range 35°N to 40°N . (Corrected for quasi-seasonal effects).

40°N to 45°N . While the daytime values for the two latitude ranges are virtually identical, the electron density contours show a quite different behavior during the rest of the day. (The fact that there are no data between 2200 and 2400 LMT is the result of poor performance of the topside sounder during these night hours as the result of leakage of terrestrial noise which causes blocking of the receiver (Atkins and Chapman, 1963)). In comparing the behavior of the mean diurnal variation at the two latitude ranges, which can be termed in the geographical sense as *mid-latitudes*, one has to keep in mind, that magnetically this latitude range at the 75°W meridian is already representative of rather high dip angles;

e.g., at 39°N , 75°W , the magnetic dip is $I = 70^\circ$. It appears that this dip represents somewhat of a boundary between "normal" and "auroral" type of ionosphere. This is definitely evidenced in the behavior of spread F (Calvert and Schmid, 1964) and thus is possibly also indicative of the role of corpuscular effects in the ionosphere (cf. Mariani, 1963). Similar conclusions have been drawn on the basis of a comprehensive study of the behavior of the "bottomside" ionosphere (Wright, 1962). The different behavior of the topside in the two latitude ranges, which are only 5° apart, yet include the "dip-boundary" can best be illustrated by comparing the amplitude of the diurnal variation at selected fixed heights at the 35°N to 40°N latitude range with that at 40°N to 45°N . This

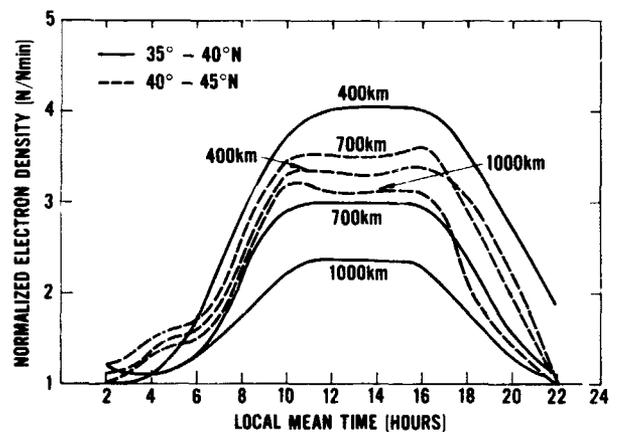


FIGURE 4.—Time variation of normalized electron density at fixed altitudes illustrating the variation with altitude of the diurnal amplitude for the two latitude ranges.

is shown in Figure 4, where the normalized electron density, i.e. the ratio of electron density at a particular local time to the minimum value of electron density at selected altitudes is plotted as a function of local mean time for both latitude ranges. It is quite obvious that the diurnal amplitude is decreasing with altitude for the latitudes 35°N to 40°N , while for 40°N to 45°N the diurnal amplitude is virtually independent of altitude. Furthermore, the absolute value of the amplitude at 400 km is higher at 35°N to 40°N than at 40°N to 45°N . The variation of the diurnal amplitude at latitudes 35°N to 40°N as a function of altitude is qualitatively what is expected from theoretical considerations (Gliddon and Kendall, 1962), how-

ever the actual decrease with altitude of the ratio of maximum to minimum density from a value of 4 at 400 km to a value of 2.3 at 1000 km is still much slower than theoretical estimates. It should be noted, however, that although the theoretical model of the F₂ region by Gliddon and Kendall is based on the solution of the time-dependent continuity equation, it is not strictly applicable to the topside ionosphere because it assumes an isothermal atmosphere whose temperature also does not show a diurnal variation. It is quite evident from Figure 4 that the diurnal behavior of the topside ionosphere, especially at latitudes greater than 40°N, is quite different from simple model concepts. The reason for the discrepancies is obviously the importance of a variable scale height as the result of diurnal variation of temperature, absence of thermal equilibrium between electrons and ions and a varying ion composition.

The above conclusion that a variable scale height is indeed responsible for the altitude behavior of the diurnal amplitude of electron density at fixed levels, is more readily apparent in "mean" vertical profiles of electron density (derived from cross sections of the electron density contours at fixed heights, as shown in Figures 2 and 3).

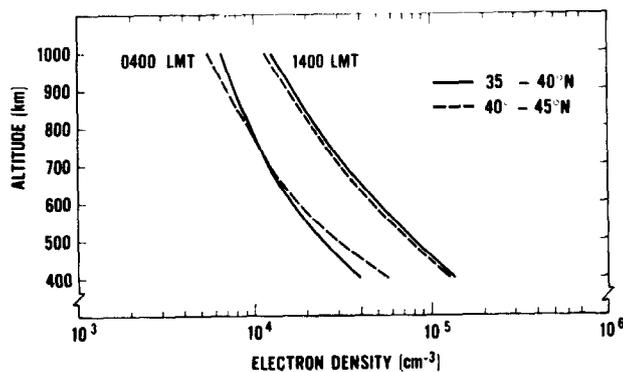


FIGURE 5.—"Mean" electron density profiles of the mid-latitude topside ionosphere corresponding to times of the diurnal temperature maximum (1400 LMT) and minimum (0400 LMT).

Figure 5 shows such "mean" profiles for the two latitude ranges at 0400 LMT and 1400 LMT, the times corresponding to the diurnal minimum and maximum of the atmospheric temperature. The problems concerning a variable scale height of the

topside ionosphere and the implications from the observations presented here will be discussed in the following section.

DERIVED QUANTITIES: SCALE HEIGHT, ELECTRON TEMPERATURE AND ION COMPOSITION

The spacing of electron density contours at fixed altitudes shown in Figures 2 and 3 is a measure of the scale height of the electron-ion gas. It is well known that in the topside ionosphere this scale height is given by

$$H' = \frac{k(T_e + T_i)}{m_+g} = \left[-\frac{1}{N} \frac{\partial N}{\partial z} \right]^{-1}$$

where *k* is Boltzmann's constant, *T_e* and *T_i* are the electron and ion temperature respectively, *m₊* is the mean ionic mass and *g* is the acceleration of gravity. From the diurnal variation of the mean electron density contours it is thus also possible to infer the diurnal variation of the scale

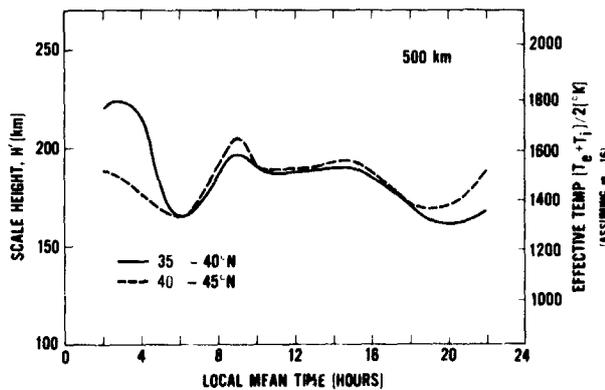


FIGURE 6.—Electron-ion scale heights at 500 km for the two mid-latitude ranges as a function of local mean time, together with corresponding scale for the effective charged particle temperature $(T_e + T_i)/2$ assuming $m_+ = 16$. The interpretation of scale heights in terms of electron temperature *T_e* and mean ionic mass *m₊* is discussed in the text.

height. Figure 6 shows the scale height *H'* at an altitude of 500 km for the two latitude ranges under consideration as a function of local mean time. In addition, a scale is given for the effective charged particle temperature $(T_e + T_i)/2$, assuming a mean ionic mass $m_+ = 16$. The latter assumption may be justifiable at the given alti-

tude, at least during the daytime. In examining the variation of H' as a function of local time one has to keep in mind that this variation is the result of variation in *electron- and ion-temperature, and ion composition*. Referring to the temperature scale, it is obvious that the effective charged particle temperature is in excess of the neutral gas temperature T_n for the given three-month period, when T_n was about $1000^\circ K$ during the day, and about $700^\circ K$ at night (cf. Nicolet, 1963). Thus, it is quite evident that thermal equilibrium is absent, at least during the day (when $m_+ = 16$ may be a reasonably good assumption). If we assume that $T_i = T_n \approx 1000^\circ$, the electron temperature T_e is of the order of $2000^\circ K$ during most of the day with an even higher value around 0900 LMT. (If m_+ is less than 16, then T_e is reduced proportionately; thus, the above value represents an *upper limit*). It is interesting to note, that actual measurements of T_e near the F2 peak for the same latitude range obtained from Blossom Point telemetry read-outs of the S-6 aeronomy satellite (Brace, Spencer and Dalgarno, 1964) during spring and summer of 1963, are in excellent agreement with our inferred electron temperatures. Even the maximum at 0900 LMT and the subsequent plateau in the electron temperature appear in both data. Thus, we are confident, that our assumptions are reasonable. Furthermore, there is the suggestion of a latitude gradient in the scale height during the day when we are able to infer electron temperature. The electron temperature appears to be higher at latitudes greater than $40^\circ N$ than below that, a conclusion, again in agreement with the actual measurements of electron temperature by Brace *et al* (1964). During the early morning hours (0200-0500 LMT) the electron-ion scale height H' is higher than, or at least comparable to, the daytime values. This cannot be explained in terms of temperature since there is good reason to believe that atmospheric heating is lower at night (assuming that the main heat source is solar EUV with a possibly constant, corpuscular component super imposed on it). The more plausible explanation is that the large scale height during this time period is the result of a change in the ion composition, indicating the presence of the lighter ionic constituents He^+ and H^+ . The latitudinal gradient in scale height is reversed during this time period, implying that in terms of

ion composition, the preponderance of the light ions is weighted towards lower latitudes. This inference finds support in observations of ion composition on the Ariel satellite (Bowen *et al*, 1964). If such a latitude gradient of ion composition does indeed exist even during the day, then the latitude-gradient in electron temperature would be even larger than that apparent from the daytime scale heights for the two latitude ranges. Ion composition, or rather the mean ionic mass m_+ at 500 km, can be inferred from the scale height if assumptions are made concerning the electron and ion temperatures. Assuming first, that thermal equilibrium prevails during the night, then at 0400 LMT (corresponding to the diurnal temperature minimum) $T_e = T_i = T_n \approx 700^\circ K$ for the time period under consideration. Accordingly, the mean ionic mass at 500 km is $m_+ \approx 8$ AMU at latitudes $40^\circ N$ to $45^\circ N$ and $m_+ \approx 7$ AMU at latitudes $35^\circ N$ to $40^\circ N$. The trend seems to be reversed in the late evening hours when the scale heights are greater at higher latitudes than at the lower latitudes. However, the assumption of thermal equilibrium may not hold even during the night, since electron temperatures $T_e \approx 1000^\circ K$ have been observed at nighttime (Brace *et al*, 1964) which are in excess of the estimated neutral gas temperature for the corresponding time period. In this case the mean ionic masses quoted above would be increased to $m_+ \approx 10$ AMU ($40^\circ N$ to $45^\circ N$) and $m_+ \approx 8$ AMU ($35^\circ N$ to $40^\circ N$). In any event, it is quite obvious that the light ionic constituents must become important during the night even at altitudes as low as 500 km.

It is interesting to compare the scale heights at 500 km with the scale heights at 800 km derived in the same fashion. During the night (0400 LMT) the ratio of the scale heights $H'(800)/H'(500) \approx 2.5$ at $35^\circ N$ to $40^\circ N$ and about 2.1 at $40^\circ N$ to $45^\circ N$, while during the day (1400 LMT), this ratio is about 1.9 at the lower latitudes and 1.6 at the higher latitudes. Taking into account the altitude variation of the acceleration of gravity g , and making the same assumptions concerning the electron and ion temperature as before, it is again possible to infer the mean ionic mass at 800 km. Table 1 summarizes the inferred mean ionic mass at the two latitude ranges, considered here.

TABLE 1.—*Inferred Mean Ionic Mass m_+*

Altitude	$T_e/T_{i,n}$	0400 LMT		1400 LMT	
		35°-40°N	40°-45°N	35°-40°N	40°-45°N
500 km	1 >1	~7 AMU ~8 AMU	~8 AMU ~10 AMU	(16 AMU)	(16 AMU)
800 km	1 >1	~3 AMU ~3.5 AMU	~4 AMU ~5 AMU	~6 AMU ~9 AMU	~7 AMU ~11 AMU

It is obvious from Table 1 that light ionic constituents (H_e^+ , H^+) must become important at 800 km even during the day; during the night H^+ must be an important constituent, at least at the lower latitudes. To infer the relative abundance of ionic constituents becomes somewhat hazardous, since the additional assumption would have to be made that the light ions are in diffusive equilibrium, an assumption which may not be justified even at altitudes as high as 700 km (Bauer, 1964). Thus, the mean ionic mass m_+ can only be used as an *indicator* of the presence of ions lighter than O^+ , while no reliable and unique data on the abundance of the individual light ions can be inferred from the electron-ion scale height under the present circumstances.

CONCLUSIONS

Although it is not possible by means of satellite observations to obtain directly the true diurnal variation of the topside ionosphere, a *mean* diurnal variation can be constructed with proper precautions from daily observations over an extended period of time. The mean diurnal variation of the topside ionosphere at mid-latitudes based on Alouette observations from October to December 1962 exhibits the following features:

1. There is a significant difference in the diurnal behavior between the latitude ranges 35°N to 40°N and 40°N to 45°N, the boundary of which appears to be associated with the 70° magnetic dip line.
2. The vertical cross section of the topside ionosphere is indicative of an electron scale height *variable* with altitude, latitude and time.
3. There is definite evidence of absence of thermal equilibrium, at least during the daytime, with $T_e/T_{i,n} \approx 2$ at 500 km.
4. The mean ionic mass inferred from the scale height indicates that the light ionic constituents H_e^+ and H^+ are already of importance during the day at 800 km and during the night at an altitude as low as 500 km.
5. There is some indication of opposite latitude gradients in electron temperature and mean ionic mass, the electron temperature increasing with latitude.

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