Nocturnal Ionization in the F_2 Ionospheric Region

N. C. GERSON*
Air Force Cambridge Research Laboratory, Cambridge, Massachusetts

A critical survey was made of the upper atmosphere to an altitude of 300 km, reviewing present knowledge on such factors as composition, winds, temperature, pressure and diffusion. On the basis of recent information a model atmosphere having a diurnal temperature cycle above 100 km was adopted.

Employing the adopted temperature-altitude function, it was possible to re-examine the rate of recombination occurring in the F_2 ionospheric layer. Considering a temperature-dependent recombination coefficient and including the effect of atmospheric contraction arising from the quotidian temperature variation, it was possible to reformulate the equation describing the rate of change of electron density. On this basis, the predicted value of electron density during darkness in the F_2 ionospheric layer agreed with the observed values very much better than did those obtained when the simple quadratic recombination law was employed.

In the course of the investigation, all possible sources of nocturnal ionization were examined. For the F_2 layer insufficient data are on hand to reveal the magnitude of horizontal transport phenomena, which may be of importance. Vertical transport is quantitatively treated. All other sources of ionization seem to be insufficient to maintain the observed ionization levels.

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Symbols and Constants

a = temperature lapse rate (°K/cm)
b = lapse rate of volume dissociation (cm^{-1})
e = charge on electron; base of natural logarithms
f = frequency (mc/sec.)
g = acceleration of gravity (cm/sec.^{2})
h = Planck's constant (erg sec.)
k = Boltzmann's constant (erg/mole°K)
ln = logarithm to base e
m = mass of ion (g)
mb = millibars = 1000 dynes/cm^2
n' = number of negative ions/cm^3
n = number of electrons/cm^3
p = number of positive ions/cm^3
q = rate of production of electrons by photo-ionization (cm^{-3}/sec.$^{-1}$)
r = radius of earth (cm)
s = rate of production of electrons by detachment (sec.$^{-1}$)
t = time (sec.)
z = altitude above earth (cm)
A = Avogadro's number (molecules/mole)
$H = kT/MUg$ = scale height by definition (cm)
K = degrees on Kelvin scale
M = mean molecular weight of air
N = number of neutral particles/cm^3
P = pressure (mb)
$R = R'/M_0$ = gas constant for dry air (cm^2/°K sec.^{2})
R' = universal gas constant (cm^2/°K sec.^{2})
S = intensity of radiation at the top of the atmosphere
T = temperature (°K)
U = atomic mass unit (g)
X = zenith angle of sun (radians)
α = recombination coefficient (cm^3/sec.)
θ = colatitude
ϕ = latitude
ν = frequency (cycles/sec.)
ρ = density of dry air (g/cm^3)
Ω = angular rotation of earth (radians/sec.)

Constants

$r = 6.367623\times10^8$ (cm) = mean radius of earth
$g_0 = 980.665$ cm/sec.^{2} = standard acceleration of gravity at the surface of the earth
$T_0 = 288$°K = mean temperature of earth at sea level
$P_0 = 1013.250$ mb = standard pressure at the surface of the earth
$\rho_0 = 1.2929$ (g/cm^3)** = standard density at sea level and 273°K
$M_0 = 28.979$ = mean molecular weight of air at sea level (obtained from ρ_0)
$\Omega = 7.29211\times10^{-6}$ radians/sec. = angular rotation of earth
$U = 1.6600\times10^{-24}$ (g) = atomic mass unit

I. INTRODUCTION

CONVENTIONAL theories regarding the ionosphere ascribe the production of electrons to photo-ionization by solar radiation, and the destruction

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** From M. S. Blanchard and S. F. Pickering, "A review of the literature relating to normal density of gases," Sci. Paper 529, Nat. Bur. Stand. Washington (1926).

(or removal) of free electrons by recombination or attachment to positively ionized or neutral particles, respectively. The electron density in the upper atmosphere, when considered as a function of height, exhibits four distinct maxima, the neighborhood of each being termed the D, E, F_1, and F_2 regions, in order of ascending height. These different regions, usually considered to be formed by different electromagnetic absorption processes, are observed through the reflection of radio waves radiated at ground level, and more recently by the decrease in intensity of radio waves transmitted from rockets.

The two main strata of atmospheric ionization are the E region at about 110 km and the F region, from about 200–300 km. The F region during daylight splits into F_1 and F_2 layers which merge to form one layer after sunset. In each region the ionization density reveals characteristic diurnal and seasonal variations of considerable magnitude. Although present theories adequately explain these variations in the lower regions, ionization density fluctuations in the F_2 layer appear to be erratic and not in conformance with present hypotheses. The E region varies normally with the sun's elevation and while the F_1 region varies roughly in the same fashion, the F_2 layer shows no such simple correspondence.

Ionospheric layer formations are found to be closely linked with the diurnal variations of electron density. When extra-terrestrial electromagnetic radiation is absorbed by an atmospheric constituent whose concentration decreases exponentially with increasing height above the earth's surface, the rate of absorption evinces a sharp maximum at a given level. The cause lies in the fact that absorption at any point is proportional to both the gas density and the radiation density at the point in question. If the former diminishes rapidly with increasing height and the latter with decreasing height, a sharp maximum in the rate of absorption will arise. When the absorptive process is photo-ionization (or photo-dissociation), levels of marked maxima in the rate of electron (or atom) production occur. However, because of the variation of the rate of disappearance (or rate of association) with height, the level of maximum production will not necessarily be identical with the level of maximum electron (or atom) density.

If the source of ionizing radiation is removed, as during darkness, the height of the layer of maximum electron concentration and the electron density itself will both be determined by the rate of removal of electrons. The rate of destruction proposed by the present theory for the F_2 layer should practically dissipate this region prior to sunrise.

This paper, after first examining current knowledge on the ionosphere, attempts to explain nocturnal ionization in the F_2 layer. The explanation is sought by examining the nature of the electron combination processes. For a better knowledge of the various reactions, the constitution, the pressure and the temperature in the upper atmosphere must also be investigated.

II. ATMOSPHERIC COMPOSITION AND STRUCTURE TO 300 KM

A. Composition

The earth's atmosphere consists of a mixture of gases (and suspended solid particles) retained by the earth through gravitational attraction. In the troposphere the principal constituents, in proportion by volume, are molecular nitrogen, 78 percent; molecular oxygen, 21 percent; and argon, 1 percent.[1] In addition, water vapor in highly variable quantities and several other gaseous elements and compounds are also present.

Above the tropopause the existence of ozone in a well-defined band from about 10–30 km has been shown by several researchers, the maximum volumetric percentage being close to a level of 30 km.[2–4] The investigations of the vertical ozone distribution indicate that the maximum lies slightly below 30 km, and that there is a seasonal variation in the ozone content.[5] A somewhat lower altitude was found by balloon flights which revealed a maximum ozone concentration close to 20 km.[6] Double layers have been reported in a number of instances. Götz first showed that for a high total amount of ozone, two maxima occur over middle Europe.[7] The same was also found by Tönsberg and Olsen for Northern Norway.[7a] Recent rocket flights also revealed the double layer over southern U. S., one maximum occurring at 17 km, and the second at 25 km.[8] Rocket data also indicated that above 30 km the measured concentration decreases sharply, becoming vanishingly small at 50 km. A trace of ozone was observed, however, at 67 km.

The evidence that molecular oxygen dissociates to the atomic state seems unmistakable. Rakshit[9] found that dissociation of oxygen occurs between 100–110 km, while Wulf and Deming[10] conclude that molecular oxygen dissociates to the atomic state in a somewhat narrow zone from about 94–100 km, and that above this level only atomic oxygen occurs. The findings of the latter researchers appear to be the more reliable, especially in the light of recent findings by Penndorf[11]

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and others. The presence of atomic oxygen at great heights has been amply verified by observations on the emission spectra of the polar and non-polar aurora.[12–16]

That atomic nitrogen exists at high altitudes is now generally accepted. Accumulating evidence gives strong support to the presence of some atomic nitrogen at the altitudes under consideration,[17–21] but the degree of dissociation is not known with any certainty. Contrarily, others contend that no atomic nitrogen is present in the high atmosphere.[22]

The non-observance of the spectral lines of the noble gases has been interpreted as proof of their absence in the upper atmosphere. However, it appears more probable that because of the high energies required for excitation, only an insignificant number of line spectra are emitted from these gases and that these spectra are too weak to be observed. It might be of interest at this point to remark that in addition to these elements water vapor[23] and methane have been noted in the solar absorption spectrum. The night sky spectrum also contains the lines of neutral atomic sodium,[24,25] molecular nitrogen,[26–28] and other compounds, some not yet identified. Absorption bands also support the existence of nitrogen-oxygen compounds.[29]

B. Mixing

Fairly strong evidence indicates that the lower and upper atmospheres are well mixed, probably to extreme heights. Meteorological observations show that winds, convection and turbulence provide a generally homogeneous troposphere. In support of this conclusion the volumetric percentage of oxygen has been found to be constant to greater than 25 km,[30] and of helium, to about 20 km.[31] Wind velocities at 30 km averaged at least 40 km/hr. over England.[32] From circulation patterns expected at these heights, it seems certain that winds of at least this velocity would be found at other geographic localities. Hoffmeister,[33] observing "Leuchstreifen" at an altitude of about 120 km over Germany determined that their movement indicated a fairly constant SSW–SW flow at 180 km/hr. in summer (May–September). During winter, two "Leuchstreifen" systems occurred; a SW flow at 230 km/hr. and a northerly flow at 320 km/hr., both of which varied appreciably in intensity and direction. In southwest Africa northerly and westerly flows were found, with the former more predominant.

From observed distortions of meteor trails in the altitude range 40–110 km, winds and turbulence of 80 km/hr. and higher have been indicated.[34,34a] Studies on the movement of clouds of sporadic E ionization in the height range 100–110 km, confirmed wind movements of 15–125 km/hr. over Germany[35] and 290–380 km/hr. over the United States.[36,37] The latter velocities appear somewhat high although not unreasonably so. Direct and indirect wind velocity observations above 110 km are lacking.

Early investigators once propounded a diffusive equilibrium of the atmospheric constituents (mainly N_2 and O_2) resulting in a solely hydrogen or helium atmosphere from 200 km upward.[38,39] Other workers indicated the possibility of these lighter elements remaining on the outer fringe of the atmosphere.[40,41] Little confirmation of this view may be gained from spectrographic evidence since, because of the high ionization potential of He, only a small probability exists of this element becoming ionized. It seems probable that hydrogen and some helium do escape from the earth's gravitational field. With regard to the height of diffusive equilibrium, later determinations indicated that with an upper atmosphere composed of molecular nitrogen and atomic oxygen, and having a positive temperature lapse rate, diffusive equilibrium would occur at an altitude greater than 350 km.[42]

In view of the above evidence and since arguments will be later introduced for accepting a temperature increase with height (above 100 km), it seems logical to accept complete mixing of the atmospheric constituents to at least the level of 350 km.

It might be noted that in an attempt to explain the existence of the sodium D line in the spectrum of the light of the night sky, the compound NaO (atomic weight—39) has been suggested.[43] If this hypothesis is correct, there remains no reason why argon (atomic

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weight—40) should be restricted from the upper atmosphere.

C. Temperature

The vertical temperature distribution has been considered by a multitude of investigators; it will suffice here to outline briefly the major findings and to indicate the temperature-altitude relationship to be adopted.

Several experimental and theoretical fields have contributed to the knowledge of upper air temperatures and pressures; namely:

a. Direct soundings by means of balloons and rockets.
b. Observations of compressional wave propagation.
c. Theory of atmospheric tides.
d. Spectrographic observations on the light of the night sky and the aurora.
e. Observations on meteor trails.
f. Theories of atmospheric radiative equilibrium.
g. Theory of low pressure ionic recombination.

If the upper atmosphere were in a state of thermodynamic equilibrium, then it would be anticipated that these various methods for obtaining temperatures would lead to the same value. As, however, the existence of the spectral lines is an indication that a Maxwell velocity distribution of the particles is violated, it may be anticipated that strict thermal equilibrium is not found in the high atmosphere and that the several methods may yield results biased by the experiment or the assumptions involved.[44–46]

Before arriving at any definite conclusion, this question will be discussed more fully in the following pages.

Direct measurements reveal that mean tropospheric and lower stratospheric temperatures and pressures[47,48] at least in middle latitudes, are closely given by the U. S.[49] or the ICAN[50] standard atmospheres. The U. S. standard atmosphere proposes a surface temperature of 288°K which decreases linearly with altitude to 218°K at slightly below 11 km. Above this level an isothermal stratosphere of 218°K is considered. The latter value is in general agreement with observations on the brightness of the twilight sky which indicate a constant temperature near 220°K from about 20 to 55 km.[51]

From observations on the anomalous propagation of sound waves (arising from explosions), Whipple,[52,53] Duckert,[54] and Gutenberg,[55] predicted that above the isothermal stratum temperatures should rise to a maximum value in the neighborhood of 50–60 km altitude. Such high temperatures were in accord with the theories of Taylor,[56] who studied the propagation of compressional waves formed by the explosion of the great Siberian Meteor of 1908 and the eruption of the volcanic island of Krakatoa in 1883. Pekeris[57] further strengthened the belief in a high temperature region at 60 km when he studied the semidiurnal barometric pressure cycle. To account for the two free periods of atmospheric oscillation, Pekeris found it necessary to postulate a temperature decrease above 60 km. The hypothesis was strengthened by Humphreys'[58] suggestion that the noctilucent clouds occasionally observed at 82 km were composed of ice crystals at temperatures possibly below 200°K. Direct evidence of a negative temperature lapse rate above 60 km has been obtained from rocket flights,[58a] under the assumption of a given upper atmospheric composition.

From sound wave propagation, a daytime temperature of about 370°K near 60 km is indicated. With the atmospheric tide theory a temperature about 350°K at 60 km and about 240°K near 80 km is needed. Recent rocket data reveal a temperature about 300°K at 60 km and about 260°K at 80 km; however, these data were described as being of not very great accuracy.

On the basis of meteor observations Whipple[59] concluded that a temperature of approximately 250°K should be found from 80 km to 95 km, with the data favoring a high temperature zone near or slightly above 60 km. He noted that the value of 250°K may be considered as a minimum value, and that the data cannot be taken to exclude a temperature minimum near 82 km. In the vicinity of an altitude of 100 km, a second temperature increase is required to account for the energy distribution observed in the nitrogen bands of the auroral spectrum. At 110 km, Vegard[60] reported a temperature of 218°K but Rosseland and Steensholt[61] indicated a value of 347°K. The various data given above have been plotted in Fig. 1, together with points from the NACA upper atmosphere tentative minimum, standard and maximum temperature.[62] Although the temperatures obtained by different methods and at different latitudes are not strictly comparable, it is seen that most of the data are in fair agreement.

Above 100–110 km the evaluation of temperature has produced widely varying and sometimes discordant results. On the basis of radiative equilibrium for various

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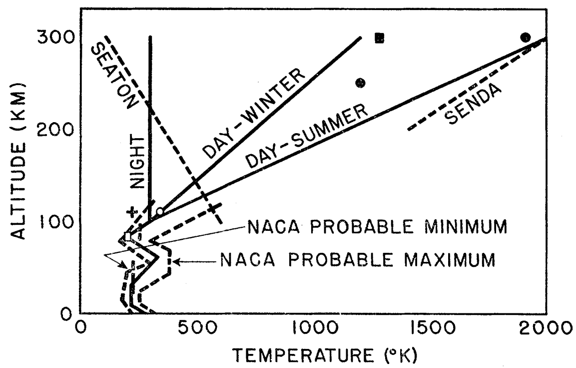

FIG. 1. Adopted temperature—altitude relationship to 300 km.

possible percentages of ozone, oxygen and water vapor in the high atmosphere, Godfrey and Price[63] found that the greatest possible daylight temperature was 3300°K, while the minimum night temperature was about 230°K. Inasmuch as these values were computed on the basis of an atmosphere saturated with water vapor above the tropopause and containing ozone to extreme heights, the temperatures given are regarded as doubtful. Similarly, doubt is attached to the values of the absorption coefficients for water vapor and ozone employed in the calculations.

Seaton[64] on the basis of ionic recombination, found temperatures at great heights which varied from about 610°K (at 100 km) to somewhat less than 100°K (at 300 km).[65, 66] It is interesting to note that Seaton in marked distinction to other investigators, found in some cases higher temperatures at 100 km than at 300 km.

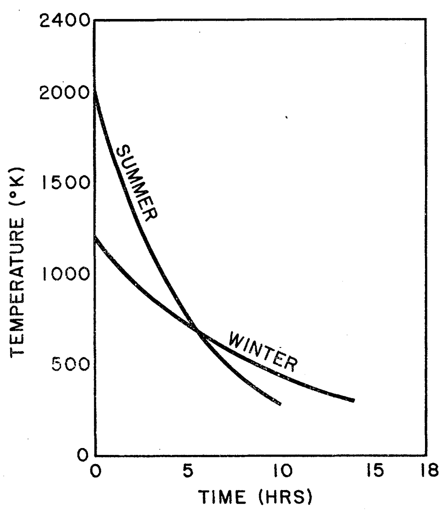

FIG. 2. Temperature variation from 2 hours before sunset until sunrise under the adopted conditions at 300 km.

In order to determine a diurnal temperature change in the upper atmosphere, the various data were carefully examined. Although some daily variation undoubtedly exists from the base of the stratosphere to 100 km, little information regarding the magnitude of the change has been found in the literature. Gowan[67] found a change of 25–30°K at an altitude of 40–50 km. Martyn and Pulley,[68] however, indicate a single temperature altitude relation to 100 km and a diurnal change above this level. In this paper a similar attitude will be followed, and the relationship adopted to 100 km will be considered to hold both day and night.

Above 100 km, a linear temperature variation to 2000°K during summer daylight and to 1200°K during winter daylight will be chosen at 300 km. During darkness it will be assumed that the temperature at each point decreases exponentially, falling to a minimum value of 300°K, just before sunrise. The net effect of this assumption gives an isothermal atmosphere at 300°K from 100 to 300 km just before sunrise. See Fig. 2.

These assumptions are in accord with other investigators. Senda[69] postulated a temperature range of 1400–2000°K between 200 and 300 km. Martyn and Pulley give a maximum daylight value of 1200°K at 250 km, and by extension, 1920°K at 300 km. Appleton finds a temperature of 1300°K in the F ionospheric region. Fuchs[70] calculated temperatures of about 1950°K at the height of maximum electron concentration in the F_2 layer. These values are somewhat higher than those reported by Babcock,[71] who estimated temperatures of 900°K by measuring the width of the auroral green line. On the other hand, Petrie,[72] examining four OII lines, concluded that the temperature varies from 3000°K to 6000°K.

On the basis of the preceding remarks, a maximum temperature of 2000°K, remaining constant from about two hours after sunrise to two hours before sunset is very possible. Similarly a temperature minimum of 300°K, occurring before sunrise is not unreasonable.

From remarks previously given regarding the absence of thermal equilibrium in the high atmosphere, even the rough agreement of temperatures obtained by such a variety of methods is remarkable. It would hardly be expected, of course, that under non-equilibrium conditions the temperatures obtained from OII spectra are the same as those determined by other means. Each determination of temperature may be considered as a measure of the energy associated with the particular process involved. Summarizing the results given above, one fact remains evident: temperatures in the upper atmosphere are high, probably near 2000°K. Further-

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more, although equilibrium is not established, the results are not so discordant that to a first approximation a Maxwellian velocity distribution of the particles may not be assumed.

On the basis of the foregoing considerations a constant temperature T_m is adopted as occurring at 300 km from after sunrise to before sunset. During the interval from two hours prior to sunset until sunrise an exponential temperature decrease is assumed, *viz.*,

$$T=T_m e^{-2ft} \qquad (2\text{–}1)$$

where

f = constant to be determined on the basis of T_s/T_m

and

T_s = presunrise temperature = 300°K.

For winter, t was taken as 14×3600 seconds and $T_m=1200$°K; for summer, $t=10\times3600$ seconds and $T_m=2000$°K.

D. Pressure

In order to determine the number of particles/cm^3 in the F_2 region of the ionosphere, a knowledge of the pressure and density at great heights is necessary. The latter quantities were determined from the hypsometric equation which was solved by employing a temperature-altitude relationship based upon the data listed above. The adopted temperature-altitude function is shown by the solid curve in Fig. 1. It assumes a linear variation of temperature with altitude, and between 94 and 100 km, a linear dissociation with altitude of molecular to atomic oxygen. The adopted temperature-altitude function passes through the points given in Table I. The major composition of the atmosphere was assumed to be free of water vapor and unchanged to great altitudes.

Determination of the pressure at each of the above altitudes was accomplished in the familiar fashion. Assuming a static atmosphere rotating with the earth, and employing the hydrostatic equation $P=-g\rho dz$ and the perfect gas equation $P=\rho RT(M_0/M)$, there results

$$dP/P=-(gM/RTM_0)dz. \qquad (2\text{–}2)$$

Before integration of (2–2) may be attempted, it must be remembered that, in general, $g=g(z)$, $T=T(z)$ and $M=M(z)$. Thus, assuming that the acceleration of gravity varies inversely with the square of the distance from the center of the earth, and that centrifugal force further detracts from its value,

$$g=g_a[r/(r+z)]^2-(r+z)\Omega^2\cos\phi \qquad (2\text{–}3)$$

where $g_a=g_0+r\Omega^2\cos\phi$.

A linear variation of temperature with altitude was also assumed,

$$T=T_1+a(z-z_1) \qquad (2\text{–}4)$$

where

T_1 = temperature at a datum level z_1.

The ratio M/M_0 in the region of dissociation may be easily shown to be

$$M/M_0=[1+b(z-z_1)]^{-1}. \qquad (2\text{–}5)$$

The appropriate values of a and b employed are easily ascertained from Table I. Substituting (2–3) to (2–5) in (2–2) we obtain

$$\begin{aligned} dP/P=-\{&(g_ar^2/R)/[T_1+a(z-z_1)][1+b(z-z_1)]\\ &\times(r+z)^2-(\Omega^2\cos\phi/R)(r+z)/\\ &[T_1+a(z-z_1)][1+b(z-z_1)]\}dz. \qquad (2\text{–}6)\end{aligned}$$

Equation (2–6) was integrated under three conditions:

(a) when $a=b=0$ as in region 2–3 (see Table I),
(b) when $b=0$ as in regions 1–2, 3–4, 4–5, 5–6, 7–8

and

(c) when $a>0$ and $b>0$, as in region 6–7.

Integration thus provides three equations for use under the given conditions, namely,

Condition (a):

$$P_3=P_2e^{A+B} \qquad (2\text{–}7)$$

where

$$A=(g_ar^2/RT)(z_3-z_2)/(r+z_2)(r+z_3),$$
$$B=(\Omega^2\cos\phi/RT)[r(z_2-z_3)+(z_2^2-z_3^2)/2].$$

TABLE I. Data on atmosphere at different altitudes.

Region	Altitude cm $\times10^5$	Temperature °K	Pressure mb	No. particles* cm^{-3}	Density† g/cm^3	Mean free path** cm
1	0	288	1013.25	2.549×10^{19}	1.226×10^{-3}	9.812×10^{-6}
2	10.76923	218	2.349×10^{2}	7.805×10^{18}	3.756×10^{-4}	3.204×10^{-5}
3	32	218	8.605	2.860×10^{17}	1.376×10^{-5}	8.745×10^{-4}
4	62	330	2.040×10^{-1}	4.479×10^{15}	2.155×10^{-7}	5.584×10^{-2}
5	84	200	1.203×10^{-2}	4.358×10^{14}	2.097×10^{-8}	5.739×10^{-1}
6	94	262.5	2.831×10^{-3}	7.815×10^{13}	3.760×10^{-9}	3.200
7	100	300	1.525×10^{-3}	3.682×10^{13}	1.771×10^{-9}	6.793
		300 (night)	3.109×10^{-11}	7.509×10^{5}	2.986×10^{-17}	3.331×10^{8}
		500	1.884×10^{-9}	2.730×10^{7}	1.086×10^{-15}	9.159×10^{6}
8	300	800	4.316×10^{-8}	3.909×10^{8}	1.554×10^{-14}	6.399×10^{5}
		1200 (day-winter)	4.028×10^{-7}	2.432×10^{9}	9.670×10^{-14}	1.028×10^{4}
		2000 (day-summer)	3.845×10^{-6}	1.393×10^{10}	5.539×10^{-13}	1.795×10^{3}

* From $N=P/kT$.
† From $\rho=PM/M_0RT$.
** From $\lambda^{-1}=\sqrt{2}\pi Nd^2$; $d=3\times10^{-8}$ cm.

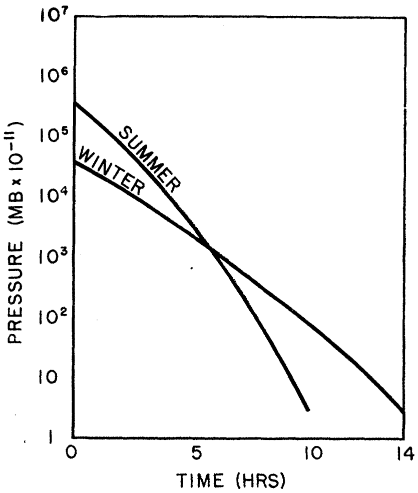

FIG. 3. Pressure variation from 2 hours before sunset until sunrise under the adopted conditions at 300 km.

Condition (b):

$$P_2=P_1K_1e^{C+D} \tag{2-8}$$

where

$K_1=\alpha^m\beta^n$,
$\alpha=(T_1/T_2)$,
$\beta=(r+z_2)/(r+z_1)$,
$m=-(g_ar^2/a_1R)[T_1/a_1-(r+z_1)]^{-2} -(\Omega^2\cos\phi/a_1R)[T_1/a_1-(r+z_1)]$,
$n=-(g_ar^2/a_1R)[T_1/a_1-(r+z_1)]^{-2}$,
$C=(g_ar^2/a_1R)(z_2-z_1)/(r+z_1)(r+z_2)[T_1/a_1-(r+z_1)]$,
$D=-(\Omega^2\cos\phi/a_1R)(z_2-z_1)$.

Condition (c):

$$P_7=P_6K_6e^E \tag{2-9}$$

where

$K_6=\alpha^p\beta^q\gamma^s$,
$\gamma=1+b(z_7-z_6)$,
$p=-(g_ar^2/R)/(a_6-bT_6)[T_6/a_6-(r+z_6)]^2 +[\Omega^2\cos\phi/(a_6R-bRT_6)][1/b-(r+z_6)]$,
$q=-(g_ar^2/R)[(a_6+bT_6)-2a_6b(r+z_6)] \times[T_6-a_6(r+z_6)]^{-2}\cdot[1-b(r+z_6)]^{-2}$,
$s=-(g_ar^2/R)/(a_6-bT_6)[1/b-(r+z_6)]^2 +[\Omega^2\cos\phi/(a_6R-bRT_6)][(T_6/a_6)-(r+z_6)]$,
$E=-(g_ar^2/R)(z_7-z_6)/(r+z_6)(r+z_7)[T_6-a_6(r+z_6)] \cdot[1-b(r+z_6)]$.

By use of the appropriate equations, pressures at each of the levels shown in Table I were calculated. The number of particles/cm³ and the mean free path were also determined.

In order to evaluate the contribution to the total barometric pressure caused by an increase in the number of particles through ionization, use may be made of the perfect gas equation $P=\rho R'T/M$. Since $N=A\rho/M$, the perfect gas equation may be written

$$P=kNT. \tag{2-10}$$

However, if only single ionization occurs, and if no electrons are lost by attachment to form negative ions $n=p$, and the pressure is given by

$$P'=k(N+n)T. \tag{2-11}$$

From Table I it may be seen that at 300 km N varies from 10^6 to 10^{10} under the adopted assumptions. In the F_2 region of the ionosphere, however, N remains at approximately $10^{10}/\mathrm{cm}^3$. As the concentration of electrons is only about 10^6–10^7, it is evident that the increased pressure brought about by single ionization is negligible. Thus, $P=P'$ effectively.

The values of pressure, mean free path, number of particles, etc., are in close agreement with those computed by Grimminger[73] and Harang.[74]

The pressure decrease during darkness at the level of 300 km for both winter and summer is shown in Fig. 3. It should be noted that the pressure variation with time at this altitude may be represented by

$$P=P_me^{-2fgt} \tag{2-12}$$

where

$P_m=P(T_m)$ and g is an empirical constant.

The value of g is easily determined to be:

Points	Value of g: Summer	Winter
1–2	3.8	3.6
2–3	4.7	4.4
3–4	5.6	5.3
4–5	6.8	

III. PHOTO-CHEMICAL PROCESSES INCLUDING PHOTO-IONIZATION

A. General Absorption

The fundamental theory regarding the formation of the ionosphere, first propounded by Chapman[75] in 1931, is based upon several rather idealized conditions. These conditions specify:

a. A uniform beam of monochromatic radiation falling from a sun in space upon a rotating earth.
b. An isothermal atmosphere where the density decreases exponentially with height.
c. Complete absorption of the radiation solely by the atmosphere.
d. An absorption of radiation proportional at any point to the atmospheric density and the radiational intensity.
e. A utilization of a constant fraction (up to 100 percent) of the absorbed radiation in dissociating or ionizing an atmospheric constituent.
f. A recombination of the products of dissociation (or ionization) exclusively with one another to

[73] C. Grimminger, "Analysis of Temperature, Pressure and Density of the Atmosphere Extending to Extreme Altitudes," Rand Corporation, Santa Monica, California (February, 1948).
[74] L. Harang, Terr. Mag. **51**, 353 (1948).
[75] S. Chapman, Proc. Phys. Soc. London **43**, 26 (1931); **43**, 483 (1931).

reform their parent particle, in accordance with the quadratic law of recombination.

g. No diffusion of the products of dissociation (or ionization) from the element of volume wherein they were formed.

The rate of the assumed earth's rotation and the rate of the assumed sun's change in declination were taken to be identical with that found for the actual earth and sun.

The rates of insolation, and of photo-ionization or photo-dissociation, were determined as functions of height, time of day, season of year and latitude. Similarly, the concentration of ionized particles also was obtained as a function of the above variables.

The final relevant equation for the absorption of monochromatic solar radiation derived under the above circumstances was

$$q(t) = BpS\rho' \exp - [z/H + B\rho' He^{-z/H} \sec X] \quad (3\text{–}1)$$

where

$q(t)$ = rate of production of ions,
B = coefficient of absorption,
p = number of ions produced by the absorption of a unit quantity of radiation,
S = intensity of radiation at the top of the atmosphere,
ρ' = an assumed mean density of the atmosphere,
z = altitude above the surface of the earth,
$H = kT/M_0 gU$ = scale height,
(H = 8.4 km when T = 300°K (with Chapman's constants)),
U = atomic mass unit,
$X = \theta + D$ = sun's zenith distance,
θ = colatitude of place of observation,
D = maximum declination of sun.

Obviously the rate of production of ions has a maximum value when

$$z = H \ln(B\rho' H \sec X). \quad (3\text{–}2)$$

Equation (3–2) gives the observed E and F_1 region ionization densities and the ionization variations observed throughout the day.

It should be noted, however, that (3–2) has been based upon rather broad assumptions which, in many cases, conflict with the atmospheric model employed for other purposes. For example, in a preceding section the pressure and density of the atmosphere at great heights were computed on the basis of more recent data, and a daylight temperature considerably higher than 300°K was employed. From the discussion on pressure it is obvious that the atmospheric density does not decrease exponentially with altitude but in a much more complicated manner. The ionizing quanta of solar radiation are not completely absorbed by the upper atmosphere (some penetrate to the ozone layer), nor does a constant fraction of absorbed radiation invariably result in dissociation or ionization. Diffusion and winds, which are also neglected, may be of appreciable importance. Nevertheless, the remarkable fact remains that the existence of the E layer is well explained and the existence of the F_1 layer is roughly clarified by Chapman's hypotheses; the observed F_2 layer does not follow this law.

B. Reactions

In order to perceive more clearly the processes of absorption involved in the upper atmosphere, the individual reactions of the particular constituents must be examined in more detail. A knowledge of these reactions is also required in order that the mathematical recombination equations may be adequately understood. Some conception of the probabilities of the various reactions may be gained from Table II, which gives the ionization and dissociation potentials for various atmospheric constituents, and the number of such quanta available from the sun radiating as a black body at 6800°K. The probability of photo-ionization decreases markedly above 10–11 electron volts, becoming extremely small in the neighborhood of 20 electron volts. Elements or compounds having an ionization potential greater than this amount, therefore, have been excluded from Table II.

The several absorption processes may be described as collision reactions (i.e., collisions between particles, or between a particle and a quantum of energy), which may be conveniently categorized in three classes:

a. dissociation and association,
b. ionization and recombination,
c. attachment and detachment.

The composition of the atmosphere to 300 km has been demonstrated to be practically identical with that found at the surface. Thus a nitrogen-oxygen-argon atmosphere is accepted, the concentration of the remaining constituents being sufficiently small to neglect them. The latter condition involves the implicit as-

TABLE II. Ionization and dissociation potentials of major atmospheric constituents.

Element or compound	Ionization potential*‡ (ev)	Dissociation potential** (ev)	Number of quanta emitted†
Na	5.12		3.5×10^{20}
NO	9.5	6.49	1.3×10^{18}
NO_2	11.0		1.65×10^{17}
NH_3	11.2		1.25×10^{17}
O_2	12.5	5.08	1.94×10^{16}
H_2O	12.56		
N_2O	12.9		1.06×10^{16}
O	13.55		4.8×10^{15}
N	14.98		4.9×10^{14}
N_2	15.51	9.76	
A	15.68		1.7×10^{14}

* First ionization level.
† With energies greater than the given ionization potential, considering sun as black body at 6800°K.
‡ C. D. Hodgmans, *Handbook of Chemistry and Physics, 1863* (Chemical Rubber Publishing Company, Cleveland, 1943).
** A. G. Gaydon, *Dissociative Energies and Spectra of Diatomic Molecules* (John Wiley and Sons, Inc., New York, 1947).

sumption that the disregarded constituents do not affect the rates of production, recombination, attachment, or detachment. As it has already been established that the presence of even small traces of a contaminating gas can markedly affect low pressure gas discharges,[76] the latter supposition is open to question.

In common with all chemical and atomic transformations, a dynamic equilibrium exists in the processes described below. Whether the reaction proceeds to the right or to the left depends upon the conditions involved. Generally, the equations proceed to the right during daylight and to the left during darkness. In the notation which follows the symbol "G" will be employed to denote a third body.

It is not the purpose of this section to enter into a full discussion of all the conceivable reactions involved, nor into complete detail regarding the probabilities of the various processes. This section merely aims to describe the reactions most expected and to discard those evidently not possible.

a. Dissociation and Association

The basic reactions are:

$$O_2+h\nu \leftrightarrows O+O \qquad N_2+h\nu \leftrightarrows N+N, \tag{3–3}$$

$$O_2+G \leftrightarrows O+O+G \qquad N_2+G \leftrightarrows N+N+G. \tag{3–4}$$

These dissociative transformations lead to a supply of atoms and determine the ratio of atomic to molecular components at any given level. Because available evidence appears to indicate little atomic nitrogen in the high atmosphere, equilibrium for the equations involving nitrogen undoubtedly occurs with very low concentration of N. Although the dissociation of oxygen proceeds only during daylight, Wulf and Deming[10] circumstantiated that the atoms thus produced will remain in the atomic state for considerable periods of time, up to perhaps several days. For this reason large scale association to reform molecular oxygen during darkness has not been considered. In this study, there has been adopted an atmosphere containing entirely molecular oxygen below 94 km, a constant gradient of volume dissociation from 94 km to 100 km, and atomic oxygen above 100 km. These limits are closely those determined by Wulf and Deming.

The forward dissociation-by-impact reaction, Eq. (3–4) requires a particle G having adequate energy to dissociate the molecule by collision; the reverse action necessitates a three-body collision. In the first instance, under the conjecture that the particles are in thermal equilibrium at temperatures below 2500°K, the probability of G having a kinetic energy greater than the dissociation potential of O_2 or N_2 is small. In the second case, because of the very low density of the F_2 region, the probability of a three-body collision is also quite small. Both reactions, therefore, are considered to be extremely infrequent. Note that if G is in an excited state the probability of (3–4) increases considerably.

b. Ionization and Recombination

$$O+h\nu \rightarrow O^+ + e \qquad N+h\nu \rightarrow N^+ + e, \tag{3–5}$$

$$O_2+h\nu \rightarrow O^+ + O^- \qquad N_2+h\nu \rightarrow N^+ + N^-, \tag{3–6}$$

$$NO+h\nu \rightarrow NO^+ + e \qquad NO_2+h\nu \rightarrow NO_2^+ + e$$

$$N_2O+h\nu \rightarrow N_2O^+ + e \tag{3–7}$$

$$O_2+G \rightarrow O^+ + O^- + G \qquad N_2+G \rightarrow N^+ + N^- + G \tag{3–8}$$

$$O+G \rightarrow O^+ + G + e \qquad N+G \rightarrow N^+ + G + e. \tag{3–9}$$

The ionization reactions (3–5), (3–6), and (3–7) are effectively restricted to daylight when ionizing quanta in great numbers are available. Because of relatively low kinetic energies of the particles, ionization by collision, Eqs. (3–8) and (3–9), is non-existent or at most negligible, except when the particle G is in an excited state. However, they undoubtedly become important factors during meteoric showers when G represents a meteor arriving from outer space. Reactions (3–6) and (3–8), requiring the presence of molecular oxygen are restricted to the region below 100 km.

The reactions listed above, together with transfer reactions and several others involving the minor atmospheric constituents, supply all the electrons, negative ions, and positive ions which collectively form the ionosphere. Inasmuch as the sun is regarded as a black body radiating at 6800°K, the most important reactions will be those requiring the lowest energy quanta (see Table II).

Recombination, the reverse of ionization, naturally occurs when the reactions of Eqs. (3–5) to (3–9) proceed to the left.

An additional reaction which may be of significance, is that which raises molecular oxygen to its excited (O_2') state:

$$O_2^+ + e \rightarrow O_2' \rightarrow O' + O''. \tag{3–10}$$

The excitation process, (3–10), occurs only if the electron has exactly the quantity of energy which is required to raise the molecular oxygen ion to an excited state, O_2'. Since the dissociation energy of molecular oxygen is smaller than its ionization energy, the excited molecular state is unstable, and dissociation eventuates. As the number of free electrons in the ionosphere having this requisite narrow energy range is small, the probability of (3–10) is correspondingly small. Furthermore, since the excitation reaction requires the presence of molecular oxygen ions, it need not be considered in connection with the F_2 layer.

The three-body collisions representing the reverse actions (3–8) and (3–9) are, as noted above, very

[76] V. J. Francis, and H. G. Jenkins, Rept. Prog. Phys. **7**, 230 (1941).

infrequent. On the basis of quantum-mechanical considerations, Bates and his co-workers[77] established that the coefficient of radiative recombination (3–5) is very small. However, because of the low values of the recombination coefficient found experimentally for the F_2 region, ionic recombination of oxygen, (3–5) may nevertheless be important.

c. Attachment and Detachment.

$$O+e\rightarrow O^-+h\nu \quad N+e\rightarrow N^-+h\nu, \qquad (3\text{–}11)$$

$$O_2+e\rightarrow O_2+h\nu \quad N_2+e\rightarrow N_2^-+h\nu, \qquad (3\text{–}12)$$

$$O_2+e\rightarrow O+O^-. \qquad (3\text{–}13)$$

The electron affinity of O is 3.08[78] electron volts,[79] and of O_2, 0.3 electron volt. As nitrogen has no affinity for electrons,[80] the attachment reactions involving both N and N_2 may be immediately disregarded. Bates and Massey[81] found that under certain conditions resonance effects may cause the reaction (3–11) to become important. In general, the possibilities of the processes (3–11) and (3–12) are similar.[82] As dissociative attachment (3–13) requires at least a 3 volt electron,[83] the probability of its occurrence in the ionosphere is remote.

The reverse reactions (3–11) to (3–13) result in detachment with the liberation of free electrons. Supplemental processes, such as

$$N_2+O^-\rightarrow N_2O+e \quad N+O_2^-\rightarrow NO_2+e, \qquad (3\text{–}14)$$

$$N+O^-\rightarrow NO+e, \qquad (3\text{–}15)$$

may also be considered, but from the remarks given above, their probability is concluded to be small.

From the previous discussion of photo-chemical processes, it is clear that free electrons may originate from the actions of ionization and detachment. The greatest contributing factors in each case are quanta of energy sufficient to remove electrons, collision reactions being for the most part ineffective. During darkness, the electron population will disappear by the reverse processes of recombination and attachment. The rate of removal of electrons called for by present theory, however, is such that the F_2 layer of the ionosphere should dissipate at a greater rate than is experimentally found. The following section, therefore, examines further actions which may be effective in producing free electrons.

[77] Bates, Buckingham, Massey, and Unwin, Proc. Roy. Soc. **A170**, 322 (1939).

[78] The electron affinity of atomic oxygen as given in the literature ranges from 1.0 to 3.8 electron volts.

[79] D. T. Vier, and J. E. Mayer, J. Chem. Phys. **12**, 28 (1944).

[80] F. L. Arnot (Nature **138**, 162 (1936)) reported observing N^-; but O. Tuxen (Zeits. f. Physik **103**, 463 (1936)) in a similar experiment could not detect any negative ions.

[81] D. R. Bates, and H. S. W. Massey, Phil. Trans. **A239**, 269 (1943).

[82] D. R. Bates, and H. S. W. Massey, Proc. Roy. Soc. **A187**, 17 (1946).

[83] H. D. Hagstrum, and J. T. Tate, Phys. Rev. **59**, 354 (1941).

IV. MAINTENANCE OF NOCTURNAL IONIZATION LEVELS

A. General

The conceivable non-solar causations of ionization in the ionosphere and particularly in the F_2 region, may be grouped in three general classes. The first class consists of possible sources of ionization, i.e., all operative methods which create ions from neutral particles. The second category comprises those actions which transfer ions from one region to another. The third subdivision embraces all procedures which tend to store or hoard ions

a. Generation of ions
   1. Radiation from interstellar space (except cosmic rays)
   2. Cosmic rays
   3. Secondary radiation from the lunar surface
   4. Particle collision
   5. Intercepted cosmic dust
   6. High speed neutral or charged particles
   7. Terrestrial radioactivity
   8. Thunderstorm activity
   9. Meteorological effects (other than thunderstorms)
   10. Meteoric bombardment

b. Transportation of ions
   1. General high atmospheric circulation
   2. Molecular diffusion
   3. Subsidence
   4. Thermal contraction or expansion
   5. Turbulence

c. Storage of ions
   1. Recombination processes
   2. Attachment processes
   3. Detachment processes

B. Generation of Ions

Although the various non-solar sources of ionization appear to be numerous, they will be found generally small and insignificant except for one factor, meteoric bombardment. For the sake of completeness, however, a brief discussion of each source is given.

1. Radiation from Interstellar Space (except Cosmic Rays)

The total number of ions produced by a given flux of radiation depends not only upon the quanta available, but also upon the flux intensity. Relative strengths of solar, lunar, and stellar radiation as given by Hulbert[84] are in the ratio of 10^6, 1, and 10^{-3}, respectively. Thus the efficiency in producing radiation of the latter source compared to the former is somewhat better than 10^{-9}. Although the surface temperatures of the great majority of stars are considerably greater than that of the sun, nevertheless the much larger distances involved reduce

[84] E. O. Hulbert, Phys. Rev. **31**, 1018 (1928).

the number of high energy quanta to very small amounts. Stellar radiation can scarcely have any measurable effect upon the nocturnal ionization level of the ionosphere.

2. Cosmic Rays

The amount of ionization effected by cosmic rays at sea level, is 1.4 ion pairs/cm^3 sec. With an increase in altitude, the number of ions produced by cosmic radiation increases to a maximum of about 400 ion pairs/cm^3 sec. near 17 km altitude (in middle latitudes), and then decreases. Johnson,[85] studying the contribution of cosmic radiation to the ionization of the upper atmosphere, concluded that only about 0.9 ion pair/cm^3 sec. are formed at altitudes of 40 to 100 km. It has been suggested[86] that cosmic-ray showers may be responsible for sporadic clouds of ionized particles; however, the number of clouds formed through cascade action is insufficient to account for the observed sporadic E regions. Because of the rarity of the upper atmosphere, cascade showers could scarcely be responsible for appreciable ionization in the higher layers. It is therefore unlikely that any significant fraction of F_2 ionization could be ascribed to cosmic radiations.

3. Secondary Radiations from the Lunar Surface

It has been proposed that radiation from the moon may possibly cause ionization in the high atmosphere. Since the temperature of the moon is in the neighborhood of 300°K, primary blackbody radiation from the lunar surface is entirely ineffective for ion production.

A second avenue of attack, however, was introduced by Stetson.[87] In correlating changes in the electron density of the ionosphere with phases of the moon, he propounded secondary radiation from the moon's surface as a causative agent in producing ionization. The secondary radiation naturally would arise on that portion of the moon illuminated by sunlight. When viewed from a given terrestrial point, the phase of the moon, and therefore the area from which secondary radiation originates, changes daily. For this reason, the effect of the moon on an ionospheric region would have not a diurnal, but a lunar monthly period. The proposed theory by itself could not account for the steady nocturnal ionization densities observed in the F_2 region.

4. Particle Collision

With the upper atmosphere in the vicinity of the F_2 layer at a maximum temperature of less than 2500°K, practically no particles exist with a kinetic energy sufficient to ionize other neutral particles by collision. During daylight, few high energy quanta are emitted from the sun radiating as a blackbody at 6800°K. However, the high energy quanta which are radiated may eject from atmospheric molecules high speed electrons which then can ionize other particles by collision. Furthermore, smaller energy quanta may raise a neutral particle to an excited state, and on collision with a second particle, ionization may occur. These reactions, however, have their greatest probability during sunlight; during darkness their effect is negligible.

5. Intercepted Cosmic Dust

In its general movement through the heavens, the earth's atmosphere encounters clouds of cosmic dust which consist of planetary, meteoric, stellar, etc., disintegration products. In attempting to explain patches of ionization which suddenly appeared in the vicinity of the E region, Dieminger, Goubau, and Zenneck[88] hypothesized ionized regions formed by cometary dust clouds. However, it has not been fully demonstrated that cosmic dust, ionized or otherwise, may be considered as a source of F_2 layer ionization.

6. High Speed Neutral or Charged Particles

It has been postulated on several occasions that corpuscles originating in the sun could act as ionizing agents for the various ionospheric layers.[89–91] However, after an extensive survey of various ionospheric-eclipse observations, Alpert and Gorozhankin[92] concluded that evidence of a corpuscular eclipse had not been established. Also, high speed neutral or charged particles emitted from the sun would hardly be expected to invest the dark portion of the earth with undiminished intensity throughout the night, even under the influence of the earth's magnetic field. For example, the diurnal auroral variation exhibits a maximum in the evening.[93] Activity apparently is strongest during evening and weaker in the morning. A similar effect would be anticipated if high speed particles streamed into the F_2 layer at lower latitudes. However, as the magnitude of nocturnal ionization in the F_2 layer remains almost constant, high speed corpuscles from the sun do not, by themselves, appear to be capable of explaining the observed electron concentration.

7. Terrestrial Radioactivity

The greatest number of ion pairs produced from this source is at or below the surface of the earth. The amount of ionization produced decreases rapidly with

[85] T. H. Johnson, Trans. Am. Geophys. Union, **16** Am. Meeting I, 35 (1935).
[86] P. M. S. Blackett and A. C. B. Lovell, Proc. Roy. Soc. **A177**, 183 (1941).
[87] H. T. Stetson, Terr. Mag. **49**, No. 1 (1944).
[88] Dieminger, Goubau, and Zenneck, Hoch: tech. u. elek: akus. **44**, 2 (1934).
[89] E. V. Appleton, and S. Chapman, Nature **129**, 757 (1932).
[90] G. Leithauser, and S. Beckmann, Zeits. f. Physik **17**, 327 (1936).
[91] R. L. Smith-Rose, Nature **157**, 48 (1946).
[92] J. Alpert, and B. Gorozhankin, W. Eng. **22**, 394 (1945).
[93] L. Vegard, *Terrestrial Magnetism and Electricity* (McGraw-Hill Book Company, Inc., New York, 1939) Vol. 8, p. 579.

elevation, becoming entirely insignificant at altitudes well below the height of the ionosphere.

8. Thunderstorm Activity

The influence of thunderstorms upon upper-air ionization stems from two causes, (a) the ionizing effect of the ultraviolet and high frequency electromagnetic spectral components in the lightning flash itself, and (b) the induction effect exerted by charged cumuliform clouds associated with thunderstorms. It would appear, however, that the former cause is very small, even for the E layer, because of the strong absorption by the lower atmosphere and ozonosphere of wave-lengths below 1700A. Any influence of lightning upon the ionosphere therefore would arise from the induction effect, and since the E region acts as an electrostatic shield for the other regions, it probably is the only region affected.

Observed results appear to strengthen this belief. Stoffregen[94] found an increased E region electron concentration in conjunction with wide-spread (undoubtedly frontal) thunderstorm activity. On investigating the influence of thunderstorm activity on sporadic E ionization, Bahr[95] obtained a correlation coefficient of 0.5, certainly indicative of some relationship between the two phenomena. Mitra[96] mentioned that thunderstorms affect ionization in the E region, but gave no quantitative results.

While these observations are interesting, they are not immediately pertinent to the study of the maintenance of nocturnal ionization levels. It is not sufficient that thunderstorm activity affect the ionization density of the ionosphere. It is necessary, in order that they may be considered as a possible cause of nocturnal ionization, that thunderstorms occur at any given region with a uniform regularity throughout the day, or at least with uniform regularity during the night. Neither of these conditions, in general, is found to exist. Furthermore, all effects thus far found indicate E, but not F_2 layer ionization.

9. Meteorological Effects (Other than Thunderstorms)

On the whole, statistical correlations rather than direct causes are included in "Meteorological Effects." It is believed that there is some intermediate process which transmits changes in the meteorological situation to the ionospheric environment. At this time the intermediary is unknown. It should be noted that even if strong correlations were found they would not be direct causes of nocturnal F_2 layer ionization.

Several workers have correlated weather changes or climatic differences with differences in the electron concentration at one or another of the ionospheric layers. The significance and the reason for the observed dependence over various sections of the earth are not clearly understood.

Ranzi,[97] in Italy, found irregular increases in the electron concentration of the E region when cyclonic centers were located north of the observing station. Stoffregen[94] noted an increase in the E region ionization when barometric depressions were centered within a range of several hundred kilometers. Martyn,[98] in Australia, observed a direct correlation between the ionization density of the E layer at night and the surface pressure 12 to 36 hours later. A similar but smaller time lag, also for the E layer, was observed by Martyn and Pulley.[99] Kessenikh and Bulatov[100] found a seasonal effect on the F_2 level at Tomsk and Moscow which depended upon surface climatic conditions at the two cities. Additional relationships between weather and E region ionization were also found by Bazpai[101] and Gherzi,[102] the latter being able to identify air masses in China from observations made on the (daylight) ionization densities in the sporadic E or in the F regions. It is believed that similar deductions could have been made with the night ionization concentration.

From the evidence presented, there appears to be some correlation between F_2 layer ionization and ground meteorological conditions. The relationship, however, is not sufficiently strong nor are the diurnal variations sufficiently consistent to warrant attributing nocturnal F_2 ionization to nightly weather changes.

10. Meteoric Bombardment

Great impetus recently has been given to the study of ionization by meteoric bombardment because of the fact that the resulting ionization trails are easily discernible on radar oscilloscopes. The basic question as to whether sufficient meteors strike the earth daily to account for any appreciable ionization may be answered affirmatively. Watson[103] investigating the distribution of meteors in interstellar space, found that for meteors to the ninth magnitude, 10^{10} strike the earth daily. Their total energy is about 10^{18} ergs, which, if entirely expended in ionization of atmospheric constituents, could produce over 10^{23} ion pairs/sec. if one assumes an ionization potential of 20 electron volts. However, the efficiency of ionization is by no means 100 percent; indeed, it is but a tiny fraction. In developing a theory of meteors, Maris[104] found that the impact energy of a given meteor was about 10^{-10} erg, and that the meteor trail could emit soft x-rays which would cause further ionization. Meteoric light by this theory would arise

[94] W. Stoffregen, Arkiv f. Mat. Astr. o. Fys. **B30**, No. 19 (1944).
[95] J. N. Bahr, Ind. J. Phys. **13**, 253 (1939).
[96] S. K. Mitra, Nature **137**, 503 (1936).
[97] I. Ranzi, Nature **130**, 545 (1932).
[98] F. Martyn, Nature **133**, 294 (1934).
[99] F. Martyn, and O. Pulley, Radio Res. Board of Australia, Council for Sci. and Ind. Res., Bull. 101, Rep. 11 (1936).
[100] V. N. Kessenikh, and H. D. Bulatov, Comptes Rendus, U.S.S.R. **45**, 324 (1944).
[101] R. R. Bazpai, Ind. J. Res. **13**, 57 (1939).
[102] E. Gherzi, Bull. Am. Met. Soc. **27**, 114 (1936).
[103] F. Watson, Jr., "Distribution of meteoric masses in interstellar space," Annals Har. Obs. No. 105, p. 623 (1937).
[104] H. B. Maris, Terr. Mag. **34**, 309 (1929).

mainly from ionization and excitation of the encountered atmospheric particles. Pierce[105] investigating ionization during a Leonid meteor shower, obtained ionization trails at altitudes of 100–200 km, and came to the conclusion that meteors could easily ionize all particles which they encountered. Similar conclusions had been entertained earlier by Bahr,[106] who found a sixfold increase in the E region ion density during a meteor shower. In particular, a large increase in the ionization density was observed to coincide with the time of maximum meteor bombardment. The results obtained indicated solely E (and perhaps F_1) region ionization; no evidence of an increase in the density of F_2 layer ionization was found.

Similarly, Skellet,[107–109] in a series of studies extending through several years, was able to correlate E layer ionization with meteoric showers. In his first paper it was found that the ionization produced during a shower was almost 1/13 that produced by the sun, but that during normal non-shower periods the fraction decreased markedly, becoming about $1/10^5$. His estimate of the number of meteors per day is in good agreement with Watson's. Pierce[110] came to the conclusion that normal ionospheric records may indicate about 100 meteor traces daily. This figure is somewhat low in comparison with that of Appleton,[111] who claimed that 10^3 meteors/hour may be detected at any time. The difference may be attributed to the different frequencies employed in each experiment. Appleton arrived at the important conclusion that *nocturnal ionization may be almost completely attributed to meteoric bombardment.*

The subject is far from closed, and the brief outline given above covers but a small proportion of the total number of investigations which have been made or are now in progress on the influence of meteoric bombardment to upper atmospheric ionization. The exact contribution of meteoric bombardment to F_2 layer ionization is still uncertain. It is entirely possible that the residual nocturnal ionization observed in this region is that continually produced by meteors. Further studies along this line are necessary.

C. Transportation of Ions

A second general method by which the nocturnal ionic concentration may be maintained or changed is through the transportation of ions from one region to another. Electrons and ions may be carried horizontally by means of winds and general circulatory movements of the atmosphere, and vertically by means of subsidence, turbulence and thermal contraction or expansion. Diffusion may act both horizontally and vertically. Horizontal movements will be treated collectively.

In general a purely horizontal or a purely vertical movement does not occur in the atmosphere. However, the predominant motion is considered here.

1. Horizontal Transport

A general examination of transport phenomena is undoubtedly much more difficult than an investigation into the various possible sources which may give rise to nocturnal ionization. At extreme heights, a knowledge of the pressure, temperature, other essential variables and their dependence upon space and time is lacking. Although some work towards clarifying the temperature distribution at extreme heights has been accomplished by Seaton,[65,66] widespread and reliable results are sparse.

In Section II it was indicated that the atmosphere up to an altitude of 100–200 km is well mixed through the action of high winds and turbulence. Observations by Trowbridge[112] and Kahlke[113] tend to confirm this conclusion. Notwithstanding the lack of positive information at higher levels, it does not seem unreasonable to suspect that a general circulation system exists, set in motion by the greater heating of the equatorial than polar atmosphere. The unequal insolation at the equatorial and polar zones by setting up pressure and temperature gradients, requires, for equilibrium, a general circulation pattern which would move the warmer, lower latitude constituents polewards. Such a longitudinal, poleward motion would then become modified through the action of the Coriolis force, which would make the motion prevailingly west-east or east-west.

A movement of the described type would characterize the trajectories of neutral particles. However, the charged ions carried by the wind systems would describe much more complicated paths inasmuch as they would also be acted upon by the earth's magnetic field. Under the influence of the geomagnetic lines of force, the positive and negative ions would separate. A preponderance of ions of one sign would finally tend to move in a common direction, forming a lateral electric current. Michel and Burkhard[114] have postulated a similar current in an effort to clarify various F_2 phenomena. A shell current of this type may also explain some observed variations of the earth's magnetic field, and possibly the small change in F_2 layer electron density noticed during eclipses.

Although the problem is complex, it is nevertheless entirely possible that through a series of intricate steps transport phenomena may be of great importance in removing ionized particles from the sunlit regions and bringing them to the dark regions of the upper atmos-

[105] J. A. Pierce, Phys. Rev. **59**, 625 (1941).
[106] J. N. Bahr, Nature **139**, 470 (1937); Ind. J. Phys. **11**, 109 (1937).
[107] A. M. Skellet, Proc. I.R.E. **22**, 132 (1935).
[108] A. M. Skellet, Proc. I.R.E. **23**, 117 (1936).
[109] A. M. Skellet, Nature **141**, 472 (1938).
[110] J. A. Pierce, Phys. Rev. **71**, 88 (1947).
[111] E.V. Appleton, and R. Naismith, ONR, London, TROANOR 1–46 (December 1946).
[112] C. C. Trowbridge, Monthly Weather Rev. **35**, 390 (1907).
[113] S. Kahlke, Ann. Hydrog. **49**, 294 (1921).
[114] G. Michel, and Burkhard, Hoch: tech. u. Elek: akus **62**, 157 (1943).

phere. Further investigations along these lines could be undertaken profitably.

Because of the long free paths in the F_2 region, the process of diffusion is subject to some of the same restrictions found immediately above for winds; i.e., charge separation, current flow, etc. Bagge[115] has shown the expected result that when diffusion is taken into consideration, the free electron concentration decreases more rapidly than when diffusion was not active. As it is more active in the (higher particle energy) sunlit regions than in the cooler, darker regions, diffusion would undoubtedly transfer ions into the night regions of the atmosphere. At any given point such an action would probably be most vigorous a few hours after sunset and a few hours before sunrise. The observed increase in ion density prior to sunrise may very well arise from the combined action of diffusion and winds. As indicated above, however, more work along these lines is necessary before conclusions may be drawn with certainty.

2. *Vertical Transport*

Rawer[116] has suggested that subsidence may be a factor in the formation of sporadic E patches. Since a correlation has been found between surface weather conditions and F_2 region ion concentration, it is not inconceivable that subsidence may also play a role in distributing ions downward from the upper extent of the F_2 stratum into lower regions of the same (F_2) layer. This action, if sufficiently intense, may form a new maximum in the observed electron concentration, or it may merely act to strengthen the maximum already existent.

In the troposphere subsidence may transport a particle downward through an appreciable distance in a time interval ranging from hours to at most a few days. It therefore does not seem impossible for a similar effect acting through a much greater vertical distance but in about the same time to occur in the F_2 region. However, this effect is assuredly not of a nocturnal nature, but on a somewhat longer and more irregular time basis.

The mass movement of ions through atmospheric expansions or contractions was once advanced by Appleton and Naismith[117] and Hulburt[118] to explain the F_2 layer electron concentrations which are greater in January than in June. However, although this theory appeared to hold for the northern hemisphere, it did not explain the observed variations in the southern hemisphere.[119] The theory further broke down upon an examination of the temperature differences required (in the northern hemisphere) to explain the observed winter-summer ionic densities.[120] However, while they fail to interpret the seasonal changes, expansion-contraction effects nevertheless may elucidate the observed quotidian or ecliptic variations.[121]

Temperature changes in the upper atmosphere give rise, naturally, to expansion or contraction of that region as the case may be. After sunset, solar insolation is absent, and the upper atmosphere cools and begins to contract and increase in density. If we neglect for the moment the loss of electrons through attachment or recombination, the ionic population will increase in the same ratio as that of the density of all particles. In addition, because of pressure and temperature gradients now established across the sunset (or sunrise) zone, winds tend to flow from the sunlit to the dark regions. As the winds have a vertical component both horizontal and vertical transport of ions is involved. Pierce[121] has postulated an oscillatory vertical motion about a mean position. Although such an oscillation has been observed in eclipses, there is little or no evidence in the literature of it having been observed after sunset. It seems more probable, however, because of damping and other effects that such a vertical oscillation if it does exist would quickly die down.

It may be concluded, therefore, that contraction, by moving ions downward, and winds, by bringing ions across the sunset or sunrise zone, tend to increase the ion density in the dark regions of the ionosphere.

The effect of turbulence may be very difficult to predict. Small scale turbulence undoubtedly affects to a marked degree the micro-ionospheric structure. It is very probable that with existing equipment and technique, ionospheric changes less than a given magnitude cannot be detected. In general, turbulence will cause an ionospheric layer to become well-mixed and uniform, but while turbulence exists, strong discontinuities in the layer may appear. There is insufficient knowledge available at upper levels to examine this subject quantitatively.

D. Storage of Ions

In this section will be treated not only processes which tend to store electrons for quick release at a later time, but also those actions which tend to maintain ionization. Since it requires at most about $\frac{1}{4}$ or $\frac{1}{5}$ the energy needed for photo-ionization, detachment is considered a storage phenomenon. If a large number of electrons in the F_2 region become attached to atomic oxygen during daylight, the possibility of the reverse action exists, i.e., detachment operating at night to liberate electrons in quantities sufficient to maintain the observed levels of ionization. As large numbers of electrons could be stored in this manner, attachment and detachment processes must be investigated in greater detail.

[115] E. Bagge, Physik. Zeits. **44**, 163 (1943).
[116] W. Rawer, Naturwiss. **28**, 577 (1940).
[117] E. V. Appleton, and R. Naismith, Proc. Roy. Soc. **A150**, 685 (1936).
[118] E. O. Hulburt, Terr. Mag. **40**, 193 (1935).
[119] Berkner, Wells, and Seaton, Terr. Mag. **41**, 173 (1936).
[120] D. F. Martyn, and O. O. Pulley, Proc. Roy. Soc. **A154**, 455 (1936).
[121] J. A. Pierce, Proc. I.R.E. **36**, 8 (1948).

Contrarily, recombination is considered as a persistence action, since a change in its rate can suddenly absorb, liberate, or cause to be maintained, a supply of electrons. By accelerating or decelerating the rate of recombination, the electron concentration in a given ionospheric region may change appreciably in a very short time. For example, if one recombination coefficient is operative during daylight and another after darkness, it may be entirely possible to explain the rather steady F_2 region nocturnal ion density on this basis. Some justification exists for the use of different recombination coefficients during daylight and darkness, and even for taking the recombination coefficient to be variable. Strong arguments are in evidence for not regarding the recombination coefficient as a constant, but as a function of time. Under these circumstances, the time dependence must be taken into consideration before integrating the recombination equation. It should also be noted that a similar reasoning applies to the rate of attachment. Thus all three rates (recombination, attachment and detachment) may be of singular importance in explaining nocturnal F_2 ionization levels.

E. Summary

From the brief survey of processes which may be effective in maintaining nocturnal F_2 region ionization, it is clear that at least three subjects bear further investigation; (a) meteoric bombardment, (b) transportation of ions, and (c) storage of ions.

Ionization of the upper atmosphere by meteoritic impact is now being intensively studied by a number of researchers. Accumulated evidence seems to indicate clearly that a definite contribution to E layer ionization is brought about during meteor showers. However, during non-shower periods the contribution is very small, and in the F_2 layer, is negligible. Ignoring, then, item (a) it will be shown below that item (c) is probably the principal factor in maintaining nocturnal ionization densities in the F_2 ionospheric region.

V. RECOMBINATION IN THE IONOSPHERE

A. Generalized Reactions

Before proceeding with the equations and assumptions involved it is desired to define clearly the term recombination. Recombination denotes the combination of negative and positive ions present in a mass of gas to give neutral particles. The term, unfortunately, is not restricted to a single type of action, but includes in its broadest sense such general processes as initial-, preferential-, columnar-, volume-, etc., recombination. However, when dealing with very low pressures, with ions randomly distributed in space and with equal numbers of positive and negative ions, only volume recombination need be considered. Volume recombination, therefore, is usually assumed to occur in the ionosphere.

In quantitative studies the recombination process is usually described by a factor, α, known as the coefficient of recombination. Usually the coefficient defines an idealized recombination reaction which is not necessarily encountered in practice.

Assume that ideal conditions exist, and that the electrons combine only with positive ions. The well-known law of recombination then may be expressed, when ionizing radiations are present, as:

$$dn/dt=q(t)-\alpha np. \tag{5-1}$$

With a mixture of gases, Eq. (5-1) becomes the series of simultaneous equations

$$\sum_{i=1}^{r} dn_i/dt=\sum_{i=1}^{r}\left[q_i(t)-n_i\sum_j \alpha_{ij}p_j\right], \tag{5-2}$$

where the subscript i refers to the ith type of negative ion and the subscript j to the jth type of positive ion. A similar set of equations, may, of course, be written for the positive ions.

The relationships (5-2) lead to a differential equation of the rth order for each component n_i; however, under certain conditions some of the terms may be neglected, thus leading to a lower order. Fortunately, in ionospheric work, most of the terms in the equation can be shown to be insignificant, giving rise to a differential equation of the first order.

Equation (5-2) may be employed to obtain an "effective" or generalized recombination coefficient. The generalized recombination coefficient, often employed in ionospheric problems, includes more than one type of recombination reaction. It should also be noted that since the left-hand side of (5-2) gives the rate of change in density of a negative ion of the ith type, the right-hand side must implicitly involve not only recombination but also attachment.

To obtain a generalized recombination coefficient, proceed with Eq. (5-2) in the expanded form

$$\sum_i dn_i/dt=\sum_i\left[q_i(t)-n_i\alpha_{i1}p_1-n_i\sum_{j=2}\alpha_{ij}p_j\right], \tag{5-3}$$

or

$$\sum_i dn_i/dt=\sum_i\left[q_i(t)-n_iA_{i1}p_1\right], \tag{5-4}$$

where

$$A_{i1}=\alpha_{i1}\left[1+\sum_{j=2}\alpha_{ij}p_j/\alpha_{i1}p_1\right]. \tag{5-5}$$

Furthermore, if

$$L=\sum_i n_i \quad \text{and} \quad Q=\sum_i q_i$$

$$dL/dt=Q-Anp_1 \tag{5-6}$$

where

$$nA=\sum_i A_{i1}n_i. \tag{5-7}$$

A is defined as the generalized coefficient of recombination for the negative ion of the ith type.

Naturally, because of the generalization involved, the physical detail in (5–6) and (5–7) is largely obscured. As one example consider the case where, for simplicity, electrons disappear by two distinct processes (recombination and attachment, two different rates of recombination, etc.). If the second disappearance process commences after the first has ceased to operate, the true recombination curve exhibits a discontinuity between the two processes. Note, however, that A cannot describe such a hiatus. The important distinction between the recombination coefficient α and the generalized recombination coefficient A is that the former characterizes a process while the latter describes the net effect of a series of processes.

As will be shown below, Eq. (5–6) is usually employed in ionospheric investigations although (5–1) is specified. To clarify the last statement a specific case involving recombination in the ionosphere will be considered.

B. Recombination and Attachment

The ionized regions of the upper atmosphere include not only neutral atoms, neutral molecules, positive ions and electrons, but also negative ions. Since the relative effectiveness of ions in reflecting a penetrating radio wave is inversely proportional to their masses, a concentration of approximately 10^4 negative ions is necessary to cause the same reflectivity as one electron.[122] However, this number of negative ions would cause the separation between the ordinary and extraordinary reflected waves to be reduced a measurable amount—an amount which hitherto has not been observed. It may, therefore, be concluded that although some attachment occurs, the proportion of negative ions to electrons in the distinct ionospheric layers is well below $10^4/1$. Thus attachment, by removing a free electron to form a negative ion, acts in exactly the same fashion as does recombination: in both cases the electron is effectively removed and does not interact with a probing radio wave.

Because of the above ambiguity in the process involved, the rate of decay of electrons during darkness has sometimes been attributed to recombination with positive ions,[117] and at other times to attachment to neutral oxygen particles.[123,124] While the first process agrees nicely with the theory describing the formation of the ionospheric layers, the second also appears very possible because of its high rate of probability. Attempts to reconcile the conflicting views have been made by Massey and others,[125,126] who considered attachment as a storage process which released free electrons during darkness. As the rates of attachment and detachment are both high compared to the rate of recombination, these workers considered an equilibrium ratio existing between the number of electrons and the number of negative ions. A variation in the concentration of electrons and negative ions could then occur through the action of electron recombination with positive ions. Under this hypothesis, the effective process in the removal of free electrons can be attributed to recombination with positive ions. The simple recombination coefficient, however, is replaced by an effective recombination coefficient which includes electron recombination, negative ion recombination and the electron-negative ion ratio.

It should be remarked, however, that one of the greatest failings of an attachment theory arises from the fact that it can explain neither the observed ionized layers nor their variation under solar activity. The process of attachment will not result in regions where the maximum electron concentration varies throughout the day and is somewhat symmetrical about local noon. Contrarily, if electrons disappear solely by recombination, the stratification and regular variation of the ionosphere are easily explained. For example, employing expression (5–1) when $n=p$,

$$dn/dt+\alpha n^2=q(t) \tag{5–8}$$

where α may be regarded as a constant. At equilibrium $dn/dt=0$ and $n=(q/\alpha)^{\frac{1}{2}}$, i.e., the electron concentration n is a maximum where the rate of electron production $q(t)$ is a maximum. The relative importance of each mechanism has not yet been clearly resolved, and it appears best to consider both processes acting simultaneously.

The positive ions most likely to be present in the F_2 region are O^+, N_2^+ (and possibly N^+). In the following discussion it will be assumed that these ions are sufficiently similar in recombination characteristics to allow the employment of a single recombination coefficient for all three reactions. This assumption appears reasonable and considerably simplifies the resulting analysis.

At a particular ionospheric level, let N, n, n' and p be the concentrations of neutral particles, electrons, negative ions and positive ions, respectively. Neglecting diffusion[127]

$$dp/dt=q(t)-\alpha' n'p-\alpha np, \tag{5–9}$$

$$dn/dt=q(t)+s(t)n'+cn'N-\beta nN-\alpha np, \tag{5–10}$$

$$dn'/dt=\beta n-s(t)n'-cn'N-\alpha' n'p, \tag{5–11}$$

where

$q(t)$ = rate of photo-ionization of neutral particles ($\text{cm}^{-3}/\text{sec.}^{-1}$),

$s(t)$ = rate of photo-detachment of electrons from negative ions (sec.^{-1}),

[122] E. V. Appleton, Proc. Roy. Soc. **A141**, 697 (1933).
[123] T. Tukada, Report Radio Research Japan **7**, 121 (1937).
[124] F. L. Mohler, J. Research Nat. Bur. Stand. **25**, 507 (1940).
[125] H. F. W. Massey, Proc. Roy. Soc. **A163**, 542 (1937).
[126] Bates, Buckingham, Massey, and Unwin, Proc. Roy. Soc. **A170**, 322 (1939).
[127] J. Sayers, Rep. Prog. Phys. **9**, 52 (1943).

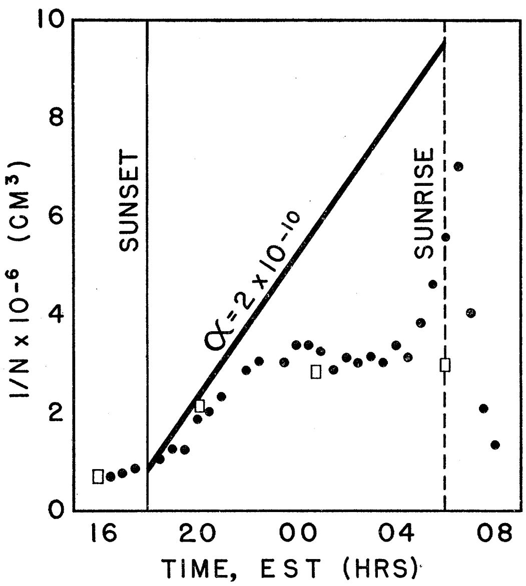

FIG. 4. Predicted and observed electron densities from 2 hours before sunset until sunrise. The solid line indicates the course of ionization expected with a constant recombination coefficient $=2\times 10^{-10}$ cm³/sec. (January 15–16, 1947).

c = rate of detachment of electrons from negative ions by collision (cm³/sec.),

β = rate of attachment of electrons to neutral particles (cm³/sec.),

α = coefficient of recombination of electrons with positive ions (cm³/sec.),

α' = coefficient of recombination of negative and positive ions (cm³/sec.).

The fact that only oxygen is available for electron attachment will be considered implicit in the attachment coefficient β. If the ionosphere is taken as electrically neutral,

$$n'+n=p. \tag{5–12}$$

Massey has indicated that the expressions βnN and $(cn'N+s(t)n')$ are each considerably greater than any of the other terms. Their ratio may therefore be considered as approximately equal to unity, whence a factor K may be defined as

$$K=n'/n=\beta N/(cN+s(t)). \tag{5–13}$$

Taking the time derivative, we have

$$dn'/dt=Kdn/dt+ndK/dt. \tag{5–14}$$

Adding Eqs. (5–11) and (5–10), and noting that $n(1+K)=p$, one obtains

$$dn/dt=q(t)/(1+K)-An^2 \tag{5–15}$$

where

$$A=\alpha+K\alpha'+n^{-1}(1+K)^{-1}(dK/dt). \tag{5–16}$$

A is a generalized coefficient of recombination. Generally, unless K is changing rapidly with time, the last term on the right will be small so that effectively $A=\alpha+K\alpha'$. At night $q(t)=0$, whence

$$dn/dt=-An^2. \tag{5–17}$$

The identity of the generalized recombination coefficient A, in Eqs. (5–6) and (5–15) should be recognized.

In the F_2 region, Wooley[128] under certain assumptions found $K=1/100$ and studies by other researchers[126, 129] also give the concentration of negative ions to electrons as small or negligible. *Thus, the net effect of both attachment and detachment processes may be ignored in the F_2 ionospheric layer.*

It is concluded therefore that the process of recombination alone exists. Recombination will be studied in more detail, in order to provide a probable explanation of nocturnal F_2 layer ionization.

C. Recombination

At very low pressures such as are found in the ionosphere, the problem of initial, preferential or columnar recombination need not be considered. The problem is truly one of pure volume recombination and, if there is no diffusion, the Thomson recombination theory is applicable. On the average, it will be assumed that in the upper atmosphere, equal numbers of ions diffuse into and out of a given volume so that diffusive equilibrium is maintained.

In Thomson's theory

$$\alpha=\pi D^2 w(\bar{u}^2+\bar{v}^2)^{\frac{1}{2}}, \tag{5–18}$$

where

$D=2e^2/3kT$ = radius of active attraction,

w = probability that an impact within the sphere D will result in recombination between the ions involved,

$\bar{u}=(3kT/m')^{\frac{1}{2}}$ = average thermal velocity of negative ions,

$\bar{v}=(3kT/m)^{\frac{1}{2}}$ = average thermal velocity of positive ions,

m' = mass of negative ion,

m = mass of positive ion.

In the Thomson theory,[130] the probability of recombination, w, is highly pressure dependent so that with decreasing pressure w and α decrease rapidly. At very low pressure where the number of particles becomes small, the probability of a three-body collision within the radius of effective attraction D may be extremely remote. Below a certain pressure, this probability falls below that arising from a direct recombination process which has been neglected in the Thomson theory.

The process of direct, neutralization recombination may ensue without the necessity of a third body when ions of opposite sign come within a critical distance d

[128] R. v.d. R. Wooley, Proc. Roy. Soc. **A187**, 403 (1946).

[129] R. Jouaust, "L'Ionosphere," Rev. d'optique, p. 77 (1946).

[130] J. J. Thomson, and G. P. Thomson, *Conduction of Electricity through Gases*, I (Cambridge University Press, London, 1928), p. 40.

of each other. In this case there are considered positive and negative ions, with an electron transfer from the latter to the former when both are separated by less than the distance d. The negative ions, although their total number is much less than that of the positive ions, thus act as catalysts in the process of neutralization recombination. In this theory, if the average radius of capture is d, the value of the recombination coefficient becomes[131]

$$\alpha = \pi d^2(\bar{u}^2 + \bar{v}^2)^{\frac{1}{2}} \qquad (5\text{–}19)$$

instead of the Thomson Eq. (5–18).

Equation (5–19) may be written in the form

$$\alpha = FT^{\frac{1}{2}} = \alpha_0 e^{-ft} \qquad (5\text{–}20)$$

where $\alpha_0 = FT_m^{\frac{1}{2}}$.

Loeb indicates that d is the average effective collisional radius when a Boltzmann velocity distribution for both types of particles exists. The value of the recombination coefficient given by this relationship is independent of pressure and depends only upon the relative ionic velocities and the average effective collisional radius.

In order to obtain one which closely represents that found experimentally, the value of the recombination coefficient will be taken from experimental data. Based upon a reasonable number of ionospheric measurements,[132–136] a value of the recombination coefficient $\alpha_0 = 2\times10^{-10}$ will be adopted for a temperature $T_m = 2000°K$. On this basis d is of the order of 10^{-8} cm, in agreement with that determined from other sources.

The nocturnal temperature decrease which was previously postulated manifests itself as an atmospheric contraction which by transferring particles downwards from upper levels increases the density at a lower level. Thus the F_2 ionospheric stratum sinks and the particle density within the layer increases as a result of the general atmospheric contraction. If the rate of change in the number of particles/cm³ is given by dN/dt, the rate of increase in the number of ions may be represented as a proportion thereof, i.e., by

$$x(t) = (n/N)(dN/dt). \qquad (5\text{–}21)$$

The term $x(t)$, negative during an expansion and positive during a contraction, may be defined as an influx function. The complete equation representing the rate of change of ionization in the ionosphere when the net effect of attachment and allied reactions may be neglected thus becomes

$$dn/dt = q(t) + x(t) - \alpha n^2. \qquad (5\text{–}22)$$

Equation (5–22) is the complete equation for electron disappearance during a contraction of the atmospheric fluid.

Before proceeding with the integration of (5–22) a few remarks are necessary regarding the value of N to be employed. As described in Section II, the hypothesized temperature and resulting pressure distributions during darkness at 300 km are given by $T = T_m e^{-2ft}$ and $P = P_m e^{-2fgt}$, respectively. Thus, the change in particle density at 300 km is given by

$$N = (P_m/kT_m)e^{-2f(g+1)t}. \qquad (5\text{–}23)$$

Equation (5–23), however, does not represent the change in density within the F_2 ionospheric region. Considering the F_2 layer to follow the isobaric surface defined during daylight by $P = P_m$, the variation in particle density during darkness in this surface is given by

$$N = (P_m/kT_m)e^{2ft}. \qquad (5\text{–}24)$$

This equation will be considered to represent the inflow of particles to the F_2 ionospheric region.

During darkness when $q(t) = 0$, Eq. (5–22) becomes

$$dn/dt = x(t) - \alpha n^2. \qquad (5\text{–}25)$$

By employing (5–20), Eq. (5–25) on quadrature gives

$$1/n = (1/n_0)e^{-2ft} + (\alpha_0/f)e^{-ft}(1 - e^{-ft}), \qquad (5\text{–}26)$$

if $t = 0$ when $n = n_0$. Note that when $T_m = \text{constant}$, $f = 0$, and $\alpha_0 = \alpha$. Under these conditions, (5–26) reverts to the equation

$$1/n - 1/n_0 = \alpha t, \qquad (5\text{–}27)$$

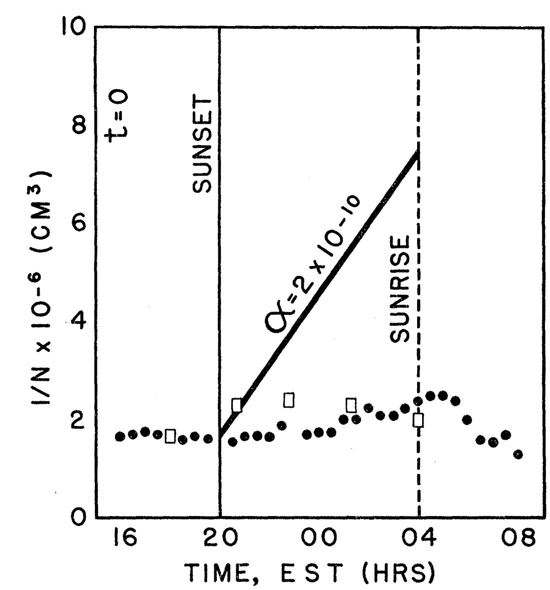

FIG. 5. Predicted and observed electron densities from 2 hours before sunset until sunrise. The solid line indicates the course of ionization expected with a constant recombination coefficient $= 2\times10^{-10}$ cm³/sec. (August 8–9, 1947).

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[131] L. B. Loeb, *Fundamental Processes of Electrical Discharge in Gases* (John Wiley & Sons, Inc., New York, 1939), p. 156.

[132] E. V. Appleton, Proc. Roy. Soc. **A162**, 451 (1937).

[133] D. R. Bates, and H. S. W. Massey, Proc. Roy. Soc. **A192**, 1 (1947).

[134] T. L. Eckersley, Nature **125**, 669 (1930).

[135] T. R. Gilliland, Am. Geophys. Union, 22 Meeting, pt. 2, 452 (1941).

[136] H. S. W. Massey, and D. R. Bates, Rep. Prog. Phys. **9**, 62 (1943).

which is identical with the result obtained when (5–8) is integrated under the condition $q(t)=0$.

VI. EXPERIMENTAL DATA

Ionospheric characteristics, including critical frequencies for the several layers, are given in publications of various governments. In the United States such information is furnished by the National Bureau of Standards, Washington, D. C. Data for Washington, D. C. are tabulated by hours of the day throughout the month.

For a study of nocturnal ionization levels during both winter and summer, observations made during the months of January and August, 1947 were chosen.[137] From the critical frequencies in the F_2 ionospheric layer, the equivalent number of free electrons/cm^3 was obtained from the familiar relationship

$$n=1.24\times 10^4 f^2 \qquad (6\text{–}1)$$

where f is the critical frequency of the ordinary wave. Although there is yet some doubt regarding the inclusion of the Lorentz' polarization term in the coefficient of (6–1), the matter need not be considered in this work.

For the comparison desired, observational data for a typical night in January and August, 1947, respectively, during which no geomagnetic activity occurred were considered. At these times, the value of $1/n$ remained relatively constant (a) throughout the night during August; and (b) within the period from about two hours after sunset to sunrise during January.

Comparisons of the results predicted by (a) a constant value of the recombination coefficient equal to 2×10^{-10} (cm^3/sec.) and (b) a temperature dependent value of the recombination coefficient together with the effect of atmospheric contraction, are shown in Figs. 4 and 5. Obviously the latter hypothesis gives results which approximate the experimental data much better than does the former.

VII. CONCLUSIONS

It is apparent that nocturnal ion densities in the F_2 ionospheric region are not clarified on the assumption of the usual quadratic law of recombination. In seeking an explanation of the greater observed than predicted electron densities, it was shown that such factors as stellar radiation, cosmic rays, secondary radiations from the moon, ionization by collision, interception of (ionized) cosmological flotsam, ionization by high energy solar ejecta, terrene radioactivity, thunderstorm activity, and meteorological induction, were negligible as sources of nocturnal ionization in the atmosphere. Meteoritic bombardment, while a substantial source of E region ionization during shower periods, appears at other times to contribute only a small amount to ionization in this region and negligibly for ionization of the F_2 ionospheric layer.

From a review of the relative magnitudes of the processes of recombination, attachment and detachment, it was concluded that the net rate of attachment-detachment phenomena during darkness is effectively zero. However, this balance easily may be one of dynamic equilibrium.

Ion density variations arising from a transportation of ions, both vertically and horizontally, are insufficiently known. In general, although data on horizontal transport phenomena are totally lacking, it is suspected that winds may carry ions fair distances from their source of production. Diffusion may be of importance because of the great molecular mean free paths at levels of 300–400 km. There is little knowledge on turbulent actions in these regions.

In an effort to predict the ion concentration measured at night a temperature decrease during darkness was assumed to exist at 300 km.

Under these circumstances an atmospheric contraction occurs which (a) lowers the altitude of the F_2 layer and (b) increases the concentration of particles within the layer. By developing this theory quantitatively, it was possible to calculate expected ionization densities which agreed with the observational data.

If the experimentally determined ionic concentrations may be attributed mainly to an atmospheric contraction, a reasonable assumption in the view of the nocturnal decrease in the F region altitude, then the described technique permits an estimate of the nocturnal temperature variation. As at the surface of the earth, the quotidian temperature ranges may vary from day to day and in this fashion may account for some of the irregularities in successive nocturnal electron densities.

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[137] CRPL, Ionospheric Data, CRPL F-30, 32; CRPL F-37, 28 Nat. Bur. Stand. Washington (1947).