CHAPTER I

INTRODUCTION

1.0 GENERAL

This thesis deals with the investigations of the D and lower E-region of the ionosphere by measuring multifrequency h.f. radio wave absorption in the ionosphere at a favourably situated tropical latitude station like Ahmedabad (23°N, 72.6°E, I = 34°N). The lower ionosphere here is free from the complexities that arise from the charged particle precipitation and the factors which produce winter anomaly observed at midlatitudes. The place is also situated near the latitude where we have the latitudinal maximum of daytime F2-ionization and most likely of radio wave absorption also. The absorption measurements provide indirectly information regarding the state of the ionosphere and can be used as a tracer for the study of the aeronomical processes in the middle atmosphere and lower ionosphere. The influences of solar activity, magnetic activity, lunar tides, meteorological disturbances etc. on the D-region dynamics can be studied. It also yields valuable data useful in practical radio communication. Since direct
in-situ measurements with rocket-borne experiments are not continuously distributed in space and time, they have only limited utility. Aeronautical studies become more meaningful if the data are available over complete periods of day, season and solar cycle. For such purpose one has to rely on ground-based radio wave propagation measurements. No regular systematic studies of absorption measurements were made at low latitude stations before 1972. Our study-programme filled up a gap with much needed information in this field. Before we proceed to give results of our studies of absorption, it would be appropriate to introduce the subject with a brief review of the present state of our knowledge of the neutral atmosphere and the D & E regions of the ionosphere. A historical account of the discovery of the ionosphere and the investigations that followed until 1950's is not included in this thesis, but they had been duly credited in the previous Ph.D. theses prepared by PATIL (1975), CHHITPA (1976) and GUPTA (1978).

1.1 NEUTRAL ATMOSPHERE

The main ionospheric regions are contained in the 60 to 600 km altitude range, but the ionosphere as a whole is related to the composition and behaviour of the neutral atmosphere throughout from ground level upwards. Atmospheric temperature, pressure, wind, production and transport of few minor species etc., exhibit considerable control over the formation of different
regions of the ionosphere and their shapes too. The information of these parameters can be found in CIRA (1972) and COBEA (1976) compilations. Fig. 1.1 shows the distribution of various constituents of the neutral gas with altitude based on exospheric temperature 1000K.

The mean CIRA (1972) gives a model for 25 to 120 km based on GROVE's data (1970) and provides a smooth transition to JACOBIA's (1971) high altitude model. Table 1.1 presents temperature, mean molecular mass and log number densities of
$N_2$, $O_2$, $O$ and $O_3$ in the altitude range 75 - 120 km. An analytical static model of temperature and composition from 20 to 2000 km altitude is proposed by ALCAIDE (1981) in which the temperature vertical profile is determined by a set of seven parameters which allows for a large variety of conditions in the stratosphere, mesosphere and thermosphere.

Table 1.1: CIRA 1972 Atmospheric model

<table>
<thead>
<tr>
<th>Ht</th>
<th>Temp.</th>
<th>Mean Log $n(N_2)$</th>
<th>Log $n(O_2)$</th>
<th>Log $n(O)$</th>
<th>Log $n(O_3)$</th>
</tr>
</thead>
<tbody>
<tr>
<td>km</td>
<td>°K</td>
<td>Mol/mass m$^{-3}$</td>
<td>m$^{-3}$</td>
<td>m$^{-3}$</td>
<td>m$^{-3}$</td>
</tr>
<tr>
<td>75</td>
<td>205.0</td>
<td>29.96</td>
<td>20.832</td>
<td>20.261</td>
<td></td>
</tr>
<tr>
<td>80</td>
<td>194.9</td>
<td>29.95</td>
<td>20.492</td>
<td>19.914</td>
<td>16.794</td>
</tr>
<tr>
<td>85</td>
<td>184.9</td>
<td>29.93</td>
<td>20.135</td>
<td>19.550</td>
<td>17.144</td>
</tr>
<tr>
<td>100</td>
<td>199.4</td>
<td>28.37</td>
<td>18.940</td>
<td>18.299</td>
<td>17.618</td>
</tr>
<tr>
<td>105</td>
<td>218.8</td>
<td>27.51</td>
<td>18.556</td>
<td>17.823</td>
<td>17.647</td>
</tr>
<tr>
<td>110</td>
<td>245.1</td>
<td>26.73</td>
<td>18.200</td>
<td>17.398</td>
<td>17.509</td>
</tr>
<tr>
<td>115</td>
<td>265.2</td>
<td>26.05</td>
<td>17.872</td>
<td>17.036</td>
<td>17.332</td>
</tr>
<tr>
<td>120</td>
<td>334.5</td>
<td>25.45</td>
<td>17.579</td>
<td>16.730</td>
<td>17.153</td>
</tr>
</tbody>
</table>

NO: Nitric oxide (NO) in the mesosphere and the lower thermosphere is an important minor constituent which is ionized by the intense solar Lyman-alpha radiation. Rocket measurements have been carried out in the mesosphere.
(NEIRA 1971) and lower thermosphere (TISONE, 1973; RUSH, 1973) and they were found to agree well. However, the fluorescent technique used in the measurement of [NO] is not accurate enough in the E-region and there is limited evidence (SECHRIST and SARCISI, 1974) that Neira's [NO] concentration should be reduced to lower limits of the error bars at heights below 90 km. Satellite measurements of [NO] fluorescence (RUSH, 1974) reveal highly variable [NO] column densities at high latitudes. Rocket measurements at Wallops Island show a substantial diurnal variation of [NO] in the 90 to 105 km region, larger concentration being present in the morning than in the evening.

OLIVER (1974) presented E-region [NO] concentration from the observations of changes in the [NO] + [O_2]/[O_2] ratio, electron density and temperature during a solar eclipse; GOLSHAN and SECHRIST (1975) derived seasonal and solar cycle variations of E-region [NO] concentration. MIRNO & SMITH (1975) determined early morning E-region [NO] from sunrise electron density measurements and found [NO] a factor of 3 to 4 times larger than obtained by Neira. They also found increases in [NO] above 120 km with increasing solar activity, in agreement with the results of GOLSHAN and SECHRIST (1975). [NO] density at auroral latitudes is generally higher than the midlatitude values (SWIDER et al, 1974) and perhaps it is higher at midlatitudes than at low latitudes (GUPTA et al, 1979; CRAVENS and STEWART, 1979). There is no asymmetry with latitude in equinoxes, but rises rapidly...

From sunrise to noon and falls only a little in the afternoon (STEWART and CRAVER, 1978).

Electron density observed by HUGHES and SMITH (1969) when compared to equilibrium calculations by SWISHA (1972) showed that Meira's [NO] profile must be reduced at least a factor of five to satisfy the ionization processes of [NO] in the initial stage of the quiet D-region. Among other estimates, the information on the level of ionization reversal with solar activity, the gradual loss of solar control in the diurnal...
variation of electron density at levels below 70 km (MITRA, 1969), the magnitude and changes in the ratio of two molecular ions \([\text{NO}^+] / [\text{O}_2^+]\) (MITRA and ROWE, 1973) and the knowledge of positive ions (HALE, 1973) etc. also lead to the estimation of mesospheric \([\text{NO}]\). Profiles obtained from different ionospheric measurements are shown in fig. 1,2. The latest measurement by BAKER et al. (1977) agrees reasonably well with those of TOHMATSU and INAGAMI (1975), and we have adopted these for our \(N(h)\) profile work.

The next most important minor constituent in the mesosphere is atomic oxygen. Concentration of \([\text{O}]\) has been estimated by several methods like, \(\text{OH}\) emission in night airglow (in-situ rocket measurement), 1.27 \(\mu\)m band of \(\text{O}_2(1\Delta g)\) and use of silver film sensor in rockets, Rutherford and Appleton laboratory fluorescence approach and simultaneous use of the mass-spectrometer with a cryo-ion source and a 5577 \(\AA\) night airglow photometer carried in a rocket over Sardinia in Spain, GOOD (1976) by \(\text{OH}\) emission in night airglow, ROGERS et al. (1973) by \(\text{OH}\) emission, DICKINSON et al. (1974) and HOLLATT et al. (1974) by fluorescence approach and EVANS et al. (1973) by 1.27 \(\mu\)m band of \(\text{O}_2(1\Delta g)\) estimated atomic oxygen concentration. RAD & MIRTY (1981) developed a new method to deduce atomic oxygen density profiles in the lower thermosphere using ground-based night airglow observations. HENDERSON (1974) used silver film sensors to measure \([\text{O}]\) between 90 and 120 km.
DONAHUE et al (1973, 1974) have deduced for heights as low as 80 km using OEO 6 satellite airglow observations on the spatial and temporal behaviour of O and discussed the implications of these results for meridional and vertical transport patterns. Reid and Chandra (1974) used OAO4 data to study the seasonal and latitudinal variation of [O], Johnson and Cottingham (1973) and Evans et al (1979) studied the meridional transport of [O] from low to high latitudes and from summer to winter hemisphere. OFFHAM (1974) reported [O]/[O2] maximum in equinoxes.

[O3]: Using Chemiluminiscence technique Hilsenrath (1971) measured [O3] in the stratosphere and mesosphere upto 70 km. Hey and Robel (1973), Robel and Hey (1974) derived nighttime [O3] in low latitude mesosphere from satellite occultation data. Measurements also exist for an intense phase of a PCA event (Hey et al, 1972) giving about 10^9 mol/cc of O3 at 60 km. All these measurements cover midlatitudes and high latitudes, no measurements are available for low latitudes. Shyamal (1981) have measured [O3] concentration profiles at Thumba with the help of rockets. The [O3] profile obtained on 17 April, 1979 gave the peak value as 3.1 x 10^{12} mol/cc occurring at 25 km and the concentration reduced to about 7 x 10^{10} mol/cc at around 78 km. The total
overhead ozone concentration at Thumba was found to be 197 D.U. (DOBSON UNIT) for the same flight. MITRA (1977) gave an equatorial model of [O$_3$] showing 10$^5$ mol/cc at 65 km to 3 x 10$^6$ mol/cc at 95 km.

$O_2(1 \Delta g)$: This is found to be quite abundant in mesosphere. EVANS et al. (1968) obtained the following equation using a radiative lifetime of 3600 sec.

$$O_2(1 \Delta g) = \mathcal{F}[O_3] \quad \ldots \quad (1.1)$$

HUFFMAN et al.'s (1971) profile of $O_2(1 \Delta g)$ is normally used for the calculation of ion-production rate. BISHOP et al. (1972) deduced $O_2(1 \Delta g)$ profile from rocket observations of the 1.27 $\mu$m emission. It varies from 2 x 10$^{10}$ mol/cc at 50 km to 8 x 10$^8$ mol/cc at 95 km in low latitudes.

$H_2O$: This is important constituent regarding the formation of heavy positive and negative ions. Observations show that a dense scattering layer is present near 80 km over the geographic pole during local summer (DONAHUE et al., 1972). It was also suggested that noctilucent clouds are weak sporadic manifestations of these persistent layers. Their suggestion that scattering layer at mesopause altitude consists of ice crystals has an important bearing on the understanding of the $H_2O$ vapour budget of the atmosphere and hence on the distribution of water cluster ions in the D-region (REID 1974). It has also been suggested
(CHESWORTH & HALE 1974) that the ice crystals may be present throughout the mesosphere over a wide range of altitudes during all seasons.

At around 70 km, water vapour gets dissociated and a substantial amount of atomic hydrogen escapes from the atmosphere. Water vapour is transported continuously upward through stratosphere and mesosphere.

1.2 RADIATIONS INVOLVED IN PHOTOCHEMISTRY

Solar ultraviolet radiations and X-rays are the principal radiations which give rise to the ionosphere. However, galactic cosmic rays and precipitating electrons are among other ionizing agents. The ionization threshold of most of the gases present in the earth's upper atmosphere corresponds to wavelengths below about 1330 Å. Table 1.2 lists these ionization potentials and their threshold wavelengths. Only \( \text{O}_2(1 \Delta_g) \), \( \text{O}_2 \), \( \text{N}_2 \), and \( \text{NO} \) have sufficiently low ionization potentials so as to be ionized by radiations at wavelengths greater than 1000 Å. Fig. 1.3(a) gives the penetration depth at which 63% of the total radiation in a given wavelength interval can reach.
Table 1.2: Ionization potentials and threshold wavelengths of ionization for atmospheric gases.

<table>
<thead>
<tr>
<th>Constituents</th>
<th>Ionization potentials $V$ (eV)</th>
<th>Threshold wavelengths $\lambda^0$ (Å)</th>
</tr>
</thead>
<tbody>
<tr>
<td>$N_2$</td>
<td>15.6</td>
<td>796</td>
</tr>
<tr>
<td>$H_2$</td>
<td>15.4</td>
<td>803</td>
</tr>
<tr>
<td>$N$</td>
<td>14.5</td>
<td>852</td>
</tr>
<tr>
<td>$CO$</td>
<td>14.0</td>
<td>885</td>
</tr>
<tr>
<td>$CO_2$</td>
<td>13.8</td>
<td>899</td>
</tr>
<tr>
<td>$O$</td>
<td>13.6</td>
<td>910</td>
</tr>
<tr>
<td>$H$</td>
<td>13.6</td>
<td>911</td>
</tr>
<tr>
<td>$OH$</td>
<td>13.3</td>
<td>949</td>
</tr>
<tr>
<td>$N_2O$</td>
<td>12.9</td>
<td>961</td>
</tr>
<tr>
<td>$O_3$</td>
<td>12.8</td>
<td>969</td>
</tr>
<tr>
<td>$H_2O$</td>
<td>12.6</td>
<td>985</td>
</tr>
<tr>
<td>$O_2$</td>
<td>12.0</td>
<td>1028</td>
</tr>
<tr>
<td>$O_2(1\Delta g)$</td>
<td>11.1</td>
<td>1118</td>
</tr>
<tr>
<td>$NO_2$</td>
<td>9.76</td>
<td>1270</td>
</tr>
<tr>
<td>$NO$</td>
<td>9.27</td>
<td>1330</td>
</tr>
</tbody>
</table>

Fig. 1.3(a): The penetration depths of solar radiation incident normally on the atmosphere. 63% of the incident energy in a given wavelength interval is lost above the altitude shown.
1.3 THE D-REGION OF THE IONOSPHERE

The D-region of the ionosphere, traditionally defined as the region of ionization below 100 km is shown to be a link between the non-ionized stratosphere below and the dense plasma above. It may be divided into three parts.

1) The upper D-region (85-90 km) which is mostly ionized by X-rays ($\gamma < 60 \text{ A}^0$) and Lyman-β ($\gamma = 1025 \text{ A}^0$).

2) The middle D-region (65-85 km) in which $[NO]$ is ionized by penetrating solar hydrogen Lyman-α radiation ($1215 \text{ A}^0$) and $O_2(\Delta g)$ by radiation of $1108 \text{ A}^0$, and

3) The lower D-region or the E-layer (below 65 km) which is ionized by galactic cosmic rays (GCR).

The ionization produced by background (non-flare) solar X-radiation in the spectral region less than $8 \text{ A}^0$ is probably the most influential agency for free electrons in the upper D-region.

In the context of ionizing radiations, references are made to the earlier works of Muller (1935), Baerer (1952), Nicolst and Aikin (1960), Chamberlain (1961), Popoff and Whitten (1962), Popoff et al. (1964), Cullane et al. (1964), Bourdeau et al. (1966). Fig.1.3(b) shows the spectral distribution of solar radiation responsible for ionizing the various species in the lower ionosphere (Bourdeau et al., 1966).
NICOL (1945) pointed out the importance of solar hydrogen Lyman-alpha (1215 Å) as a source of ionization for the lower ionosphere. Nitric oxide, [NO], though a minor constituent, exists in adequate quantity to account for the major part of the D-layer ionization. HALL, DAVIES, and HILGER (1963) estimated the total intensity of the solar Ly-alpha radiation as $5.1 \times 10^{-3}$ ergs/cm²/sec. The solar hydrogen Ly-alpha flux varies by a factor of 1.5 to 2 from solar minimum to maximum, whereas its X-ray (<8 Å) flux varies by a factor of 200 in general to 600 (during intense solar flares) as against its absorption by a factor of about 2. At solar minimum, CCR may be the dominant ionization source at altitudes up to 70 km, whereas at solar maximum, the role of CCR may not be important above 65 km.
MECOUST and AIKIN (1960) calculated a nitric oxide profile and computed its volume-ionization rate profile for the solar Lyman-alpha nitric oxide source. The observed electron density profile is in good agreement with that computed from Meira’s model of [NO] and the available data of solar Ly-α radiation, if the recombination coefficient in the lower ionosphere is fairly large. Lower values of the recombination coefficient require [NO] densities smaller than given by Meira below 87 km (TAUBNER et al., 1975). Rocket measurements also show such difference from Meira’s model (TOSHIKAWA & IWAGAMI, 1974, 1976).

1.3.1 POSITIVE IONS IN THE D-REGION

Mass spectrometer observations were made by NAGISI and BAIK (1965), NAGISI (1967), NAGISI et al. (1972), GOLDBERG and BLUMLE (1970), GOLDBERG and AIKIN (1971, 1973), KRANKOWSKY et al. (1972), ZHIBEU et al. (1973) and others. Most important feature revealed from these measurements is that the D-region should comprise of \( \text{O}_2^+ \) and \( \text{NO}^+ \), and the dominant ions below 82 km are hydrated clusters of the type \( \text{H}^+(\text{H}_2\text{O})_n \) where \( n = 1, 2, \ldots \) but \( n = 4 \) (Fig. 1.4a). Several attempts have been made to interpret these measurements through model calculations.
With reduced rocket velocity and reduced but changeable draw-in electric field, it was observed by ARNOLD et al. (1973) that the major ions below 65 km and a mass of 73⁺ corresponding to $\text{H}^+ (\text{H}_2\text{O})_1^+$, and ions of mass 55⁺, i.e., $\text{H}^+ (\text{H}_2\text{O})_3$ and 37⁺ i.e., $\text{H}^+ (\text{H}_2\text{O})_2$ were the next important. 48⁺ i.e., $\text{NO}^+ (\text{H}_2\text{O})$ ions were also observed and new ions of mass corresponding to $\text{NO}^+ \cdot \text{CO}_2$, $\text{NO}^+ \cdot \text{H}_2$ and $\text{O}_2^+ \cdot \text{CO}_2$ were found above 65 km. Various reaction schemes have been proposed to explain the formation of water cluster ions. Both $\text{NO}^+$ and $\text{O}_2^+$ are suggested to be precursors for such clustering following independent chains (FERGUSON and FREEMAN 1969, FREEMAN and FERGUSON 1969, BURKE 1970). It is believed that during normal condition, the
Fig. 1.4(b): Six-ion scheme of Mitra and Rowe

HO* chain predominates whereas during PCA, the clustering is possible through O2+ chain. CHANDRAGLY et al. (1981) found that positive ion composition measurement sometimes show the presence of molecular ions HO* and O2+ around 65 km along with hydrated protons. A simplified six-ion chemistry model with reaction rates given in Fig. 1.4(b) (MITRA and ROWE 1972) and it was extended further to include the influence of aerosols in the formation of heavier cluster ions (MITRA 1981). We shall revert to the D-region processes in sec. 1.8.
THE E-REGION

The region above 90 km and below 150 km i.e., the E-ledge is generally considered as the E-region. The radiations responsible for the formation of this region are solar radiations in the spectral region > 10 Å, Ly-β and Ly-continuum (λ < 910 Å). The problems to be resolved in this region involve measurements of photolization cross-sections, solar radiation fluxes, concentration of atmospheric species and recombination coefficient. The earliest theories of photolization of molecular oxygen by UV radiation (MITRA 1952, RAWER 1952) are referred for historical interest. As rocket measurements yielded more and more data, the emphasis shifted to the effect of X-radiation particularly in the 10-200 Å region. NICOLET (1962) summarized the most popular contemporary theories by considering the following ionization processes:

1) Ionization of molecular oxygen and molecular nitrogen by X-radiation of 31-100 Å.

2) Ionization of molecular oxygen by MW radiation in the Ly-β line (1025.7 Å) and Ly-continuum (865-912 Å), and

3) Ionization of atomic oxygen by MW radiation in the Lyman continuum (λ < 910 Å) and X-radiation.
Using the ratio of the energy spectrum distribution of 90-95 A°, 85-40 A° and 40-31 A° as 8 : 4 : 1, NICOLET (1962) obtained the profiles of production rate of electrons for the species O2 and H2. A more recent picture of electron pair production rate is shown in Fig. 1.5 taking into account the currently available estimates of radiation fluxes and ionization cross sections (GUPTA 1978).

It is evident that Ly-β radiation may be important in determining the position of the E-region peak. Atomic oxygen is ionized principally by X-radiation in the 100-110 km region; above 110 km, ionization by EUV radiation becomes dominant. For the E-region ionization, references are also made to the works of FRIEDMAN (1962), WATAHANE and HINTERMEIER (1962) and NORTON et al. (1963).

Fig. 1.5: Ion-production rate of various gas constituents in the height range 60-110 km due to different radiations for solar zenith angle 40° (after Gupta & Kotalia 1979).
1.5 LOSS MECHANISMS

The equation of continuity may be written as:

For electrons,

\[
\frac{dN}{dt} = q - \alpha_pN^+N^+ - \beta Nm + \gamma \bar{N} \bar{m} + f \bar{N}^- \]

\[ \cdots \cdots (1.2) \]

and for negative ions,

\[
\frac{dN^-}{dt} = -\alpha_i \bar{N}^+N^+ + \beta Nm - \gamma \bar{N} \bar{m} - f \bar{N}^- \]

\[ \cdots \cdots (1.3) \]

and \[N^+ = N + N^- = N (1 + \lambda)\]

\[ \cdots \cdots (1.4) \]

where \(N\) is the electron concentration, \(N^+\) positive ion concentration, \(N^-\) negative ion concentration, \(N_m\) molecular oxygen concentration, \(n\) the neutral species concentration, \(q\) electron production rate, \(\alpha_p\) average electron-ion dissociative recombination coefficient (averaged over all species of positive ions), \(\alpha_i\) ion-ion recombination coefficient, \(\beta\) attachment coefficient, \(\gamma\) negative ion collisional detachment coefficient and \(f\) negative ion photo-detachment rate. Diffusion terms are omitted because they are insignificant as compared to other terms for conditions in the lower ionosphere.
In the lower ionosphere, electrons are removed in three-body reaction which results in the formation of negative ions. The ion subsequently disappears either by recombination with a positive ion or by detachment of the electron due to collisions and radiation. At most times, attachment and detachment occur so rapidly that they nearly balance each other, so that the ratio \( \lambda \) of the concentrations of negative ions to free electrons can be calculated by equating the rates of production and loss of ions, thus

\[
\]

Hence

\[
\frac{[O_2^-]}{N} = \frac{K_1 [O_2][M]}{K_2 [M] + K_3 [O] + K_4 [I]}
\]

... ... (1.5)

When attachment and detachment of electrons occur, it is necessary to modify the continuity equation that relates the concentration of electrons to the rates at which they are produced and lost. It is supposed that the rates of attachment and detachment are not necessarily equal, and the resultant rate at which the electrons in unit volume are lost by conversion into ions will be represented by \( L \). Writing down the continuity equation for rate of change of electron density and negative ion densities and equivalently of nett positive
charge density with negative charge density, it can be shown that

\[ \frac{dN}{dt} = \frac{q}{4\pi \lambda} - (\alpha_D + \lambda \alpha_1) N^2 \ldots \ldots (1.6) \]

where \( \alpha_D \) is now the resultant recombination rate coefficient for electrons with all positive ions.

Under quasi-equilibrium condition which exists for most of the sunlit hemisphere, \( \frac{dN}{dt} = 0 \), so

\[ N^2 \approx \frac{q}{(1 + \lambda)(\alpha_D + \lambda \alpha_1)} \approx \frac{q}{\alpha_{\text{eff}}} \]

where \( \alpha_{\text{eff}} = (1 + \lambda)(\alpha_D + \lambda \alpha_1) \ldots \ldots (1.7) \)

Hence, the effective recombination coefficient serves to denote the total loss rate of electrons by the term \( \alpha_{\text{eff}} N^2 \) in the ionospheric continuity equation just as that given for the Chapman-Layer.

At heights where the number of negative ions is negligible, i.e., above 80 km, the effective recombination coefficient is determined by dissociative recombination of the various species of positive ions with electrons,

\[ \text{i.e., } \alpha_{\text{eff}} = \alpha_D \left( h > 80 \text{ km} \right) \ldots \ldots (1.8) \]
Average $\lambda_D$ is given by

$$\lambda_D = \frac{\lambda_{M_1} [M_1^+] + \lambda_{M_2} [M_2^+] + \lambda_{M_3} [M_3^+]}{\Sigma M^+} \ldots \ldots \ldots \ldots \ldots \ldots (1.9)$$

Here $[M^+]$ in the numerator within parenthesis is the number density of positive ions of masses $M_1$, $M_2$ etc in a.m.u. $\Sigma M^+$ is the total number density of positive ions (which at heights above 60 km is equal to the electron density) and $\lambda(M)$ is the dissociative recombination coefficient of the positive ion of mass $M$. Thus the effective recombination coefficient is determined primarily by the positive ion composition. Using the values of $\lambda(M)$ from SKEPI et al (1972), equation (1.9) can be written as:

$$\lambda_D = \{1.1 + 0.5 (T/300)^{-1} + 0.2 (T/300)^{-1} + 2.6 +$$

$$4.0 + 4.9 + 6.1\} \times 10^{-6} \text{ cm}^3 \text{sec}^{-1} \ldots \ldots (1.10)$$

The values in the brackets 1.1, 2.6, 4.0, 4.9, 6.1 which are temperature-independent refer to the positive ions viz. $H^+(H_2O)$, $H^+(H_2O)_2$, $NO^+(H_2O)$, $H^+(H_2O)_3$ and $NO^+(H_2O)_2$. The temperature dependent $\lambda_D$ is for $NO^+$ and $O_2^+$ in sequence.
Torr et al. (1977) gave

$$\lambda_D(\text{NO}^+) = 0.42 \times 10^{-6} \left( \frac{T}{300} \right)^{-0.85} \quad \ldots \quad (1.11)$$

Liu et al. (1973) had given the value of dissociative recombination of \( \text{O}_2^+ \) as

$$\lambda_D(\text{O}_2^+) = 0.2 \times 10^{-6} \left( \frac{T}{300} \right)^{-0.7} \quad \ldots \quad (1.12)$$

Haug & Landmark (1970) showed that the equation

$$N = \left[ \frac{\nu \lambda_{\text{eff}}}{2} \right]^{\frac{1}{2}}$$

is obeyed in the lower and upper D-region with \( \lambda_{\text{eff}} \sim 10^{-7} \text{cm}^3\text{sec}^{-1} \) for \( \text{O}_2^+ \) and \( \text{NO}^+ \)-dominated upper region and with \( 10^{-5} \text{cm}^3\text{sec}^{-1} \) for heavy water cluster ion-dominated lower region. Liu et al. (1973) measured recombination coefficients for the water cluster series at various temperatures and reported values in the range from \( 10^{-5} \) to \( 10^{-3} \text{cm}^3\text{sec}^{-1} \).

The simple molecular species, \( \text{O}_2^+ \) and \( \text{NO}^+ \), by contrast, have recombination coefficients of \( 3.5 \times 10^{-7} \) (Kashikar & Biondi, 1968) and \( 7.5 \times 10^{-7} \text{cm}^3\text{sec}^{-1} \) (Gunter & Shav, 1965) respectively at D-region temperatures. Kasikita et al. (1975) studied the electron production and loss rates in the equatorial ionosphere using a geostationary satellite signal and found the average production and loss rates equal to \( 2.8 \times 10^{14} \text{m}^{-2}\text{sec}^{-1} \) and \( 1.9 \times 10^{14} \text{m}^{-3}\text{sec}^{-1} \) respectively. They also found that both the rates show seasonal variation, and daily values of production rate shows good correlation with \( 10^{-7} \) of solar flux. The
presence of different kinds of positive ions seriously affects the effective loss rates of electrons and at present there are very few in-situ measurements, scattered in space and time, which makes modelling a difficult task.

Fig.1.6 and 1.7(a) present the profiles of $\lambda$ and $\alpha_{\text{eff}}$ respectively obtained by various workers (DERHLEN & LADELL, 1973). $\lambda$ is about 0.5 at 80 km and 0.6 at 65 km. MITRA (1975) gave theoretical values of $\lambda$ at 70 km as 0.9 under quiet solar condition, 0.55 under disturbed (solar flare) condition and 0.4 under PCA condition. CHAKRABARTI et al (1977) have shown that the enhanced electron-ion production during PCA caused by particle precipitation is not sufficient to explain the changes in parameters like $\alpha_{\text{eff}}, \lambda$ or $N$. The decrease in $\alpha_{\text{eff}}$ can only be explained by lowering the quiet-time lumped rate constant, $B$, for conversion of $\text{NO}^+ \text{to } \text{H}^+ (\text{H}_2\text{O})_n$. At around 70 km the decrease in the value of $\alpha_{\text{eff}}$ can be explained by lowering $\lambda$ and $B$ both due to decrease in $O_3$ and $O$ concentration by a factor of 2.5 and 10 respectively. Fig.1.7(b) gives the $\alpha_{\text{eff}}$ profiles at different solar zenith angles calculated by GIPTA and KOTAXA (1979) using the ratio $q/h^2$.

The electron density profile was derived from partial reflection technique, absorption measurements and ionosonde data. It may be noted that there is a transition region at 60-85 km which moves up with increasing solar zenith angle. This transition level is believed to be the border for heavy cluster ions.
Fig. 1.6: Height distribution of negative ion to electron ratio, $\lambda$ (after DREHICH and LADELL, 1973).

Fig. 1.7(a): Height distribution of effective recombination coefficient, $\alpha_{eff}$ (after DREHICH and LADELL, 1973).
Fig. 1.7(b) Effective loss rate calculated from the profiles of ion production rate and electron density for different solar zenith angles. Suffixes, I, II, and III stand for respective months. The full line curve marked m gives mean ± eff for ZA 30°-50° (after GUPTA and KOTADIA 1979).

1.6 ELECTRON DENSITY PROFILES

different solar zenith angles, to fit large number of profiles measured by rocket experiments. D-region profiles with sufficient accuracy are derived from measurements of differential absorption and/or different phase (Faraday rotation) of radio waves, propagating between ground and ascending rockets. Improved resolution is possible when Langmuir d.c. probes are flown on the same rocket along with Impedance Probe (KOTANIA et al 1974, 1977). All the ground based techniques (VLF, LF, partial reflections and wave interaction) give indirect information on N-h profiles but they involve some assumptions on the loss-rates and collision frequency.

Owing to high $\alpha_{\text{eff}}$, high neutral density, short mean free path, low electron density, direct measurements of electron density in the D-region are difficult. Experimental uncertainties add to the inconsistency of the results. Also the D-layer ionization being very sensitive to the solar activity, the day-to-day measured N-h profiles vary in shape. However, in general, the daytime profiles show two maxima, one at about 65 km of $10^2$/cc and the other at about 85 km of about $10^3$/cc (MCCULLY et al, 1967, COLE et al, 1969). A steep gradient near 85 km has been explained as due to fall in the value of $\alpha_{\text{eff}}$ (REID, 1970, 1977). Below 60 km, there are practically no free electrons but only negative ions due to high attachment rates. There is very little diurnal change in the electron density.
below 70 km (BEES, 1966). Measurements by partial reflection technique show asymmetrical diurnal variation of electron density in summer (HAUG and TIRANE, 1970), while the electron density near the level of maximum is on the average larger in winter than in summer at mid-latitudes (GRAYLOT and MAISON 1969, MIZUTAYA and SHIRKE, 1966). Electron density height (N-h) profiles have been developed by HAJAR et al (1973) under the International Reference Ionosphere (IRI) programme sponsored by the IAGI and COSPAR. Datta et al (1983) tried to test the IRI model in respect of observed A1 absorption at Ahmedabad (low latitude) for low and high solar activity, but not with much success, the absorption calculated by IRI model is found to be always higher than that observed at Ahmedabad. A similar test was made by SINGH et al (1980) at midlatitude and they found no consistency between observed and calculated absorption. We have attempted to construct N-h profiles and A1 profiles which take into account some rocket measurements, partial reflection measurements and our A1 absorption measurements (GUPTA & KOTADIA 1979). Additional data on radiation fluxes, photochemical kinetic reactions etc., are available (KLINK 1979, TORM & TORM 1979, BRATAUER & SIMON 1961, HINTEREGGER 1961, JPL Rep. 1982, WHO Rep. 1982). Uptodate information on these is collected in a monogram by DESHPANDE and METIA (1983).
1.7 **SOLAR FLARES AND ASSOCIATED PHENOMENA**

During the solar flare the enhancement in the flux of electromagnetic radiations, cosmic ray particles and solar wind cause enhancement in ionization. The following effects are observed after such eruptions:

(i) **Sudden enhancement of Atmospherics at VLF and LF (SEA).** However, at certain frequencies, e.g., frequencies below 10 MHz, this effect is observed as decrease in intensity.

(ii) **Sudden decrease in short wave field strength, called short wave fade-outs (SWF).**

(iii) **Sudden decrease in intensity of cosmic radio noise (at around 20 MHz), called sudden cosmic noise absorption (SCNA).**

(iv) **Sudden change in signal strength received from a distant LF or VLF station, called sudden enhancement in signal (SES) or sudden field anomaly (SFA).**

(v) **Sudden changes in the phase of received VLF signal (SPA).**

(vi) **Sudden frequency shift observed with highly stable frequency transmitters, originated at heights above 100 km, called sudden frequency deviation (SFD).**

During flares most significant enhancement in radiation is observed in the band 0.5 – 3 A°, 3 – 8 A° and 8 – 20 A°.
Due to very large spectral hardening of the radiation below 10 Å⁰ and increasing ionizing efficiency of the radiation with decreasing wavelength, the X-rays in these ranges control the enhancements during flares with a little time lag. It has been found that (i) the effective loss coefficient \( \alpha_{\text{eff}} \) drops rapidly with increasing X-ray flux values, \( F(1 - 3 \text{ Å}^0) \), but for large values of \( F(1 - 3 \text{ Å}^0) \), the variation slows down to near constant values. At 70 or 80 km level this occurs for \( F(1 - 3 \text{ Å}^0) \times 10^{-1} \text{ ergs cm}^{-2} \text{ sec}^{-1} \) \( (1 \text{ erg cm}^{-2} \text{ sec}^{-1} = 10^{-3} \text{ watts/m}^2) \), (ii) the decrease of \( \alpha_{\text{eff}} \) at 80 km from its preflare value to its minimum for very large flares is almost two orders of magnitude, from around \( 10^{-5} \text{ cm}^3 \text{ sec}^{-1} \) to about \( 2 \times 10^{-7} \text{ cm}^3 \text{ sec}^{-1} \). The decrease at 70 km is small, from about \( 3 \times 10^{-5} \) to \( 3 \times 10^{-6} \text{ cm}^3 \text{ sec}^{-1} \) and (iii) at 80 km, the lowest value of \( \alpha_{\text{eff}} \) is of the same order as the dissociative recombination coefficients of the \( \text{O}_2^+ \) and \( \text{NO}^+ \) ions which are also normally encountered at heights above 90 km. This indicates the dominance of \( \text{NO}^+ \) and \( \text{O}_2^+ \) in the cluster-ion region during flares.

1.8 ELEcTRON AND ION-KINETICS

The reactions of positive and negative ions of atomic and molecular species with each other, with free electrons and with neutral species are grouped together under the term 'Ion Kinetics'. The reactions of greatest interest in the ionosphere are those which determine directly and indirectly
the loss rate of free electrons as stated before in eqs:(1.2) and (1.3), unfortunately, the values of the rate constants reported for various reactions sometimes differ by more than an order of magnitude, depending upon the approach and method of analysis. Further ambiguity arises from lack of knowledge of the importance of metastable states of ions and neutral particles, both in the laboratory as well as in the ionosphere. In addition to these processes, the diffusion of charged particles can in principle, be quite significant in establishing effective local electron loss rate constants. We do not have yet much information on the reaction rate constants of ions and electrons with gases.

1.8.1 D-REGION ELECTRON-ION KINETICS

The important reactions in the D-region are as follows:

a) Production of Electrons

NO₂, though a minor constituent of the atmosphere is the major source of electrons due to its photoionization by Ly-alpha radiation. It is formed under a three-body reaction of atomic oxygen and atomic nitrogen. Atomic oxygen is available due to the dissociation of molecular oxygen by Schumann continuum radiation. Atomic nitrogen is released on dissociative recombination of NO⁺ ion and an ion-interchange processes.
\[
\begin{align*}
0 & \rightarrow 0_2 + h\nu \quad \text{---------} \quad 0 + 0 \\
\text{N} & \rightarrow \text{NO}^+ + e \quad \text{---------} \quad \text{N} + 0 \\
\text{N} & \rightarrow \text{O}^+ + \text{N}_2 \quad \text{---------} \quad \text{N} + \text{NO}^+ \\
\text{NO} & \rightarrow \text{O}^+ + \text{N} + \text{N} \quad \text{---------} \quad \text{NO} + \text{N} \quad \cdots \quad (1.13) \\
\text{NO} & \rightarrow \text{N} + \text{O}_2 \quad \text{---------} \quad \text{NO} + 0 \quad \cdots \quad (1.14) \\
\text{O} & \rightarrow \text{NO} + h\nu \quad \text{---------} \quad \text{NO}^+ + e \quad \cdots \quad (1.15)
\end{align*}
\]

Reference is also made to the work of MILLA (1968) and his review paper (1969) for nitric oxide chemistry.

b) Positive Ions

Normally \( \text{O}_2^+ \) and \( \text{NO}^+ \) ions predominate in the D-region and also in the E-region. Processes for the formation of these ions are given below:

\[
\begin{align*}
\text{NO} + h\nu & \quad \text{---------} \quad \text{NO}^+ + e \\
\text{N}_2^+ + \text{O} & \quad \text{---------} \quad \text{NO}^+ + \text{N} \quad \cdots \quad \cdots \quad (1.16) \\
\text{O}^+ + \text{NO} & \quad \text{---------} \quad \text{NO}^+ + \text{O}_2 \\
\text{N}_2^+ + \text{O}_2 & \quad \text{---------} \quad \text{O}_2^+ + \text{N}_2 \quad \cdots \quad \cdots \quad (1.17) \\
\text{O}_2 + h\nu & \quad \text{---------} \quad \text{O}_2^+ + e
\end{align*}
\]

In the D-region, however, \( \text{NO}^+ \) serves as a precursor to formation of variety of hydrated cluster ions of the form \( \text{H}^+\text{H}_2\text{O}_n \) as observed by MARGOLY & BALLOY (1965) and GOLDBERG and AILIN (1971) below 80 km. Positive ions of upto 109 a.m.u.
been observed (Johannesen et al., 1972) in addition to usual lower hydrate NO$_3^-$ (H$_2$O). At night, positive hydrated ions are small by an order of magnitude below 80 km due to nonavailability of NO$_3^-$ ions. Rocket mass spectrometer measurements of Narcist and Bailey (1965) have been challenged by Ferguson and Freienfeld (1969) and Arnold et al. (1971) in view of the difficulties in deriving realistic ion concentration from the measured currents. Fig. 1.8 gives the positive ion and electron density distribution according to photochemical-diffusive model (Hunt 1973) besides that of the cluster ions shown in Fig. 1.4.

![Fig. 1.8: Positive ion and electron density distribution profile from diurnally varying photochemical diffusive model (after Hunt 1973).](image)

c) Negative ions

These mainly exist at night below 80 km. NO$_3^-$ ions concentration is roughly 300/cc. Above 80 km, presence of atomic oxygen limits negative ions due to collisional detachment
of electrons from negative ions. Ion-ion neutralization also reduces negative ions at night below 60 km. Models of positive as well as negative ions are still incomplete. Ion-ion recombination is also another important process at altitudes of about 70-75 km during the day and 80-85 km at night. Above 85 km, the radiative process becomes more important, whereas below this altitude, the three-body process rapidly becomes dominant.

Nighttime collisional detachment of electrons (PHILPS and PACK 1961, WHITTEN and POPPOFF 1962) occurs through

\[ O_2^- + e^{-} \rightarrow O_3 + e \] (1.18)
\[ O^- + e^{-} \rightarrow O_2 + e \] (1.19)

The rate constant estimated for the reaction (1.18) is \(3.3 \times 10^{-10} \text{cm}^3\text{sec}^{-1}\) at 300K and for the reaction (1.19) it is \(1.9 \times 10^{-10} \text{cm}^3\text{sec}^{-1}\) at 300K (FERSHIELD 1967) but not yet definite. Associative processes are probably limited to altitudes over 75 km, because the O concentration decreases rapidly at lower altitudes.

ARNOLD et al (1971) found negative ions of masses 111 and 125 above 80 km but no trace of NO\(^-\), H\(_2\)O hydrate. These results differ from those of MARCHI et al (1971) and HUNT (1973) and the discrepancy is not yet explained. ARNOLD and KRAUSOWSKY (1971) have given a probable model for negative
ions but they have overestimated $NO_3^-$. A detailed and
critical review of the D-region ionic structure and the electron-
ion kinetics is given by Hunt (1973). The C-layer observed
near 70 km altitude at solar zenith angle of 85° to 95° is
probably formed by photodetachment of terminal negative ions
(Sechrist 1968, Turco and Sechrist 1972).

1.8.2. **THE E-REGION ELECTRON-ION KINETICS**

The E-region is better understood than D-region
because of more reliable measurements available on the relative
concentrations of various species, ionizing radiation fluxes
and small day-to-day variability in the ionization.

a) **Electrons**

The principal process of electron-ion production is that
of photoionization of $O_2$, $N_2$ and $O$ by the ionizing radiations
narrated earlier.

\[
\begin{align*}
O_2 + h\nu & \rightarrow O_2^+ + e \quad \cdots \cdots \quad (1.20) \\
N_2 + h\nu & \rightarrow N_2^+ + e \quad \cdots \cdots \quad (1.21) \\
O + h\nu & \rightarrow O^+ + e \quad \cdots \cdots \quad (1.22)
\end{align*}
\]

Ion-neutral collision rates are too low to contribute
much to the production of electrons. The atomic oxygen is
produced as a result of photodissociation of $O_2$. The metastable
molecules like \( O_2(1\Delta_g) \), and atoms \( O(1\,D) \) and \( O(1\,S) \) are important in the kinetics of the D-layer. The electrons are destroyed mainly by dissociative recombination process, \( \alpha \) \text{eff} = 10^{-7} \text{cm}^3\text{sec}^{-1} \), while the radiative recombination is rather weak, about \( 10^{-11} \) to \( 10^{-12} \text{cm}^3\text{sec}^{-1} \) (BATS et al 1939). The nighttime \( R \)-ionization is believed to be due to Ly-alpha and Ly-beta radiations resonantly scattered by hydrogen geocorona (SWIDER 1965; KENSHIRA et al 1970) and the significant ion is \( NO^+ \). Observations of \( NO^+ \) are in good agreement with theoretical model, though those of \( O_2^+ \) are not. \( N_2H \) profiles in the \( E \)-region obtained by rocket-borne r.f. probe and propagation experiment and by ground-based ionosondes generally agree with each other. Preliminary representative models of the ionosphere up to a height of 1000 km are prepared for a low latitude and a midlatitude by RAWAT et al (1975) under a project on 'International Reference Ionosphere' sponsored by USST and COSPAR. It gives empirical formulae and the values of several parameters involved for different ionospheric regions. A revised model of IRI was discussed in a COSPAR symposium held at Graz (Austria) in 1984.

b) \textbf{Positive Ions}

As a result of photo-ionization, there is direct production of \( O^+ \), \( O_2^+ \) and \( N_2^+ \) ions as seen in the above paragraph and these undergo the processes of charge transfer.
and ion-interchange with the neutral species.

The possible reactions are given below:

\[
\begin{align*}
\text{Rate coefficients} \\
\text{NO}^+ \text{ ions} & : \text{H}_2 + O^+ \rightarrow \text{NO}^+ + N \quad 2 \times 10^{-12} \text{cm}^3 \text{sec}^{-1} \quad \ldots (1.23) \\
& : \text{H}_2 + O \rightarrow \text{NO}^+ + N \quad 2 \times 10^{-11} \text{cm}^3 \text{sec}^{-1} \quad \ldots (1.24) \\
& : O_2 + N \rightarrow \text{NO}^+ + 0 \quad \leq 10^{-16} \text{cm}^3 \text{sec}^{-1} \quad \ldots (1.25) \\
& : O_2 + \text{NO} \rightarrow \text{NO}^+ + 0_2 \quad 6.3 \times 10^{-10} \text{cm}^3 \text{sec}^{-1} \quad \ldots (1.26) \\
\text{O}_2 \text{ ions} & : O^+ + 0_2 \rightarrow O_2^+ + 0 \quad 2 \times 10^{-11} \text{cm}^3 \text{sec}^{-1} \quad \ldots (1.27) \\
& : \text{H}_2 + 0_2 \rightarrow O_2^+ + \text{N}_2 \quad 6.4 \times 10^{-10} \text{cm}^3 \text{sec}^{-1} \quad \ldots (1.28)
\end{align*}
\]

It is seen from the above reactions that the ions are quite abundant. The ratio \( O_2^+ / \text{NO}^+ \) is near unity during day and \( \ll 1 \) during night. Above the F-layer ledge, \( O^+ \) ions dominate. The ultimate products of charge transfer or ion-interchange are \( \text{NO}^+ \) or \( O_2^+ \) which then dissociatively recombine with electrons. \( \text{N}_2^+ \) ions though largely produced, are quickly removed by charge transfer and ion-molecule reaction. Atomic \( N \) and \( O \) combine to give \( \text{NO} \). Reference is made to the works of WHITTEN and POGOFON (1964), MERTON et al (1963) and RAYES (1962). So much in general about the structure of neutral atmosphere and the lower ionosphere including ionizing radiation and ion kinetics,
1.9 THEORY OF IONOSPHERIC ABSORPTION

APPLETON (1925, 1927), GOLDSTERN (1928) and HARTREE (1929) generalized LORENZ'S (1909) theory for the propagation of electromagnetic wave through a slightly ionized medium with an arbitrary inclination of the external magnetic field. The ionized region under such condition becomes a doubly refracting dispersive medium giving two indices of refraction corresponding to what are called ordinary and extra-ordinary waves and the incident plane-polarised radio wave returns after reflection from the appropriate ionization level as elliptically polarised wave. The complex refractive index given by the classical theory, known as Appleton-Hartree (A-H) formula is

\[ n^2 = (\mu - i\chi)^2 = 1 - \frac{X}{1 - iZ - \frac{\gamma_T^2}{2(1 - X - iZ)} \pm \sqrt{\frac{\gamma_T^4}{4(1 - X - iZ)} + \gamma_L^2}} \]

*** *** (1.29)

where \( \mu \) and \( \chi \) are the real and imaginary parts of the complex refractive index \( n \),

\[ X = \frac{f_N^2}{f^2} = \frac{4\pi Ne^2}{m\omega^2} = \frac{Ne^2}{\pi mf^2} \]

\[ \gamma_T = \frac{f_B\sin\theta}{f} = \frac{B\sin\theta}{2\pi mf} ; \quad \gamma_L = \frac{f_B\cos\theta}{f} \]

\[ Z = \nu/\omega ; \quad N = \text{electron density} \text{ m}^{-3} ; \quad \nu = \text{electron collision} \]
frequency $s^{-1}$; $f_p = \text{plasma frequency} = (Ne^2/m) \frac{k}{2}$. Angular frequency of the exploring wave; $e$ and $m$ are the charge in coulombs and mass in kg of electron; $f_B = \text{electron gyromagnetic or cyclotron frequency} = 2.84 \times 10^{10} \text{ Hz}$; $B = \text{magnetic induction in Weber/m}^2$; $\chi = ck/\omega$ where $k$ is absorption coefficient/nepers/m; $c = \text{free space velocity of light in m/s}$ and $\theta$ angle between the direction of radio propagation and the magnetic field. + sign gives $n$ for ordinary wave and - sign gives $n$ for extraordinary wave.

The refractive index takes complex form because of electron collisions with the neutral molecules and atoms of the gas and to some extent ions also, a factor responsible for the absorption of radio wave during their travel through the ionosphere. The effect of increase in $\chi$ is to reduce the refractive index of the medium and to increase the absorption of radio wave.

The absorption coefficient $k$ is defined as $k = \ln (E_1/E_2)$ nepers/m where $E_1$ is the strength of the incident signal and $E_2$ the strength after the signal has traversed through a distance of one meter in the absorbing medium. Thus considering upward and downward path in the case of reflection of radio wave at vertical incidence, the total absorption can be expressed in decibels as

$$L = 2 \times 6.69 \log (E_1/E_2) \text{ dB} \ldots \ldots (1,30)$$
where the subscript \( i \) stands for 'incident' and \( r \) stands for 'received' signal through the ionosphere.

Depending on the relative magnitudes of wave frequency and plasma frequency, two types of absorption occur during the passage of the wave through the ionosphere. In case when the refractive index changes appreciably within a distance of one wavelength i.e., when the plasma frequency is close to the wave frequency the corresponding absorption is deviative. In the case when the refractive index of the medium remains nearly equal to unity or changes very little in a distance of one wavelength, the absorption suffered by the wave in such medium is mostly nondeviative.

The full expression for complex refractive index is quite complicated. It is convenient to use two approximations, one of which applies when waves are propagated close to the direction of the imposed magnetic field and the other when the propagation is nearly perpendicular to it. They are called quasi-longitudinal (QL) and quasi-transverse (QT) approximations and the conditions for which they hold are

\[
\frac{\varepsilon_r^4}{\varepsilon_L^2} \ll |(1 - x - 1z)^2| \quad \cdots \quad (QL)
\]

\[
\frac{\varepsilon_r^4}{\varepsilon_L^2} \gg |(1 - x - 1z)^2| \quad \cdots \quad (QT)
\]

Except near the equator, QL approximation is fairly
satisfactory and the expression for complex refractive index may be written as

\[ n^2 \approx 1 - \frac{X}{1 - 12 \pm |X_e|} \quad \ldots \ldots (1.31) \]

On rationalising, simplifying and separating the real and imaginary terms, we get the expression for non-deviative absorption

\[ k = \frac{2 \pi e^2}{\mu mc} \frac{N_0}{\gamma^2 + (\omega + \omega_L)^2} \quad \ldots \ldots (1.32) \]

and the corresponding absorption at vertical incidence and taking \( \omega = 1 \)

\[ L_{\omega} = 2 \times 8.69 \frac{2 \pi e^2}{mc} \int V \frac{N \gamma}{\gamma^2 + (\omega + \omega_L)^2} \quad (\text{dB}) \quad (1.33) \]

Some interesting feature can be derived from a simplified relation neglecting the magnetic field. The two cases are discussed below.

1) \( \nu \ll \omega \), as in the upper ionosphere, the limiting case for non-deviative absorption gives absorption coefficient

\[ K_{n.d} \approx \frac{2 \pi e^2}{mc} \frac{N_0 \gamma}{\omega^2} \quad \ldots \ldots (1.34) \]

and the limiting case for deviative absorption gives absorption
where $\mu$ is the real component of the phase-refractive index.

It is seen from Eq. (1.34) that the non-deviative absorption is inversely proportional to the square of the frequency, but directly to the product of electron density and collision frequency, whereas the deviative absorption varies as $\sqrt{\nu}$ and the group delay, but not as inverse frequency-squared.

2) $\nu \gg \omega$ , for LF waves reflected from the lower ionosphere, Eq. (1.33) yields

$$K_{nd} \sim \frac{2\pi e^2}{mc} \frac{N}{\nu} \quad \ldots \quad \ldots \quad (1.36)$$

i.e. when $\nu$ increases, absorption decreases. This explains the observed fact of strengthening of LF signals reflected from the lower D-region during solar flares (Bellrose 1963). Strong reflections occur due to large ionization produced by the flare radiation down to as low as 50 km altitude which results in almost infinite phase velocity and total reflection.
of the wave at such low frequencies. The phenomenon of polar cap absorption (PCA) at h.f., and v.h.f. is due to the high ionization produced in polar ionosphere by energetic protons emitted from the sun during solar flares (LEINBACH & REID 1960). According to eq.(1.33), the non-deviative absorption would be maximum at the wave frequency equal to the collision frequency.

For oblique incidence, total absorption:

\[ L_{ob} = 8.69 \int k \, ds \]

\[ = 8.69 \frac{2\pi e^2}{mc} \frac{\int N \gamma ds}{(\omega_b + \omega_c)^2} \tag{1.37} \]

where \( ds = dh \cdot \sec i \) is elemental oblique path in the ionosphere, and \( \omega_b = \omega_c \sec i \) is oblique incidence radio wave frequency. We can connect the eq.(1.33) and (1.37) according to Martyn's theorem as

\[ L_{ob} = L_{v} \cos i \tag{1.38} \]

where \( L_{v} \) represents absorption of the wave at vertical incidence on a frequency equivalent of the single-hop oblique incidence frequency for a plane ionosphere and \( i \) is angle of incidence at the lower boundary of the ionosphere.

HIGON & BENNETT (1979) has recently given a modification of the A-R formula for calculation of nondeviative
absorption as

\[ A = \frac{2k}{\omega^2} \left( \frac{1}{f} \right) \int N \, dh \]

where \( k = 8.69 \frac{2\pi e^2}{m^2} \) db same as in eq. (1.37) without any condition of restricting to quasi-longitudinal or quasi-transverse situation. The correction factor from the above modification varies with latitude from 1.00 to 1.23 to the absorption values calculated by assuming quasi-longitudinal situation. e.g. at Amadebali it is 1.23, at Singapore 1.19, at Freiburg 1.07, and at Churchill 1.01. So the latitudinal variation of \( A(f + f_L)^2 \) given by George (1971) requires to be multiplied by these correction factors.

1.10 **GENERALIZED MAGNETOTRICAL THEOREY**

The dispersion and polarisation of an electromagnetic wave propagation through slightly ionised medium are conditioned by the properties of the conductivity tensor \( \sigma \) and the dielectric tensor \( C \) of the medium defined through the equations

\[ \vec{D} = \varepsilon \vec{E} + 4\pi \vec{P} = \|c\| \vec{E} \]

and

\[ \vec{J} = \frac{\partial \vec{P}}{\partial t} = \|\sigma\| \vec{E} \]

where \( \vec{D}, \vec{P} \) and \( \vec{J} \) are the displacement, polarisation and
current vectors and \( \mathbf{E} \), the electric field of the radio wave. It can be shown that any effect of the dependence of the electron collision frequency on its velocity will change \( \delta \) and \( \epsilon \) but will not alter the form of the actual wave equation which is derived from the Maxwell's wave equations in a convenient manner. Thus, basically, it is the generalisation of \( \delta \) and \( \epsilon \) of the medium for the case \( \gamma = \gamma_\infty \) rather than the case \( \gamma \) assumed nonenergetic in the classical Adi theory. The energy dependence of collision frequency was little known until PHILPS & PACK (1959) and PHILPS (1960) established experimentally, that for slow electrons in nitrogen, the collision frequency for momentum transfer was directly proportional to electron energy.

The fundamental assumptions considered in the derivation of the generalised expression for refractive index by SES and MILLER (1960) are that the imposed electric field energy is negligible compared to the thermal energy of the electron and that its velocity distribution is Maxwellian. The final expression given by SES and MILLER is

\[
\eta^2 = \frac{c^2}{\gamma^2} = (\mu - i \chi)^2 = \frac{A + B \sin^2 \theta \pm \sqrt{B^2 \sin^4 \theta - C^2 \cos^2 \theta}}{D + E \sin^2 \theta}
\]

\[\cdots \cdots \text{(1.39)}\]
where \( \theta \) = angle between direction of propagation and magnetic field,

\[
\begin{align*}
A &= 2 \left( \frac{C}{I} + \frac{C}{III} \right) \\
B &= C_{III} \left( \frac{C}{I} + \frac{C}{III} \right) + \frac{C^2}{II} \\
C &= 2 \frac{C}{I} \frac{C}{II} \\
D &= 2 \frac{C}{I} \quad E = 2 \frac{C}{III}
\end{align*}
\]

\( C_I \), \( C_{II} \) and \( C_{III} \) are defined in terms of various constants

\( a \) ---- \( f \) which in turn are expressed in terms of wave frequency,
gyrofrequency, plasma frequency, collision frequency and Dingle's function \( C_n(x) \), \( n \) being \( 5/2 \) and \( 3/2 \) as the case may be.

They have been evaluated by BURKE & HARR (1963) and given in the
book by DAVIDSON (1966) for different values of \( (\nu_m/\omega) \).

Eqn(1.39), when the magnetic field is neglected, reduces to

\[
\eta^2 = (\mu - i\chi)^2 = 1 - \chi x^2 C_{3/2}(\chi) - i \frac{5}{2} \chi x \frac{C_{5/2}(\chi)}{\nu_m^2}
\]

where \( x = 1/\nu_m = \omega/\nu_m \). Then \( \chi \ll \omega \) for small absorption

as at high radio frequency, \( \chi \ll \mu \) and \( \mu^2 = 1 - \frac{\omega^2}{\nu_m^2} C_{3/2}(\chi) \).

The absorption coefficient turns out to be

\[
k = \frac{\omega^2}{2\mu \nu_m} \cdot \frac{5}{2} \cdot \frac{1}{\nu_m} \frac{C_{5/2}(\chi)}{\nu_m^{5/2}}
\]

In this, \( \frac{5}{2} \frac{1}{\nu_m} \frac{C_{5/2}(\chi)}{\nu_m^{5/2}} \) may be compared with the A-H formula

term \( \gamma / (\nu^2 + \omega^2) \) for absorption coefficient. It is satis-
factory to use A-H formula for calculation of absorption with
the substitution \( \gamma_{eff} = 2.5 \nu_m^2 \) at high radio frequency and
\[ \gamma_{\text{eff}} = 1.5 \gamma_m \text{ at low radio frequency instead of taking } \gamma = \nu_m \]
for all radio frequencies (THORNE & PIGGOTT 1966). \( \mu \), \( \omega \) and the product \( N \) are important in the behaviour of \( k \) with height.

1.11 COLLISION FREQUENCY

Several models of \( \gamma-h \) distribution have been proposed and used.

1) a PIGGOT and THORNE (1966)

\[ \gamma_m = 8.40 \times 10^7 p \]
where \( p \) is gas pressure in mm Hg

-b SMITH et al (1976) used a \( \gamma_m \) model to fit \( N-h \) profile with absorption data as \( \gamma_m = 6.3 \times 10^5 p \)
where \( p \) is atmospheric pressure in Newtons/m²
(1 at/m² = 10 dynes/cm²).

11) NICOLAT (1953)

\[ \gamma_{en} = 5.4 \times 10^{-10} n T^4 \]
where \( \gamma_{en} \) is electron-neutral particle collision frequency, \( n \) is neutral molecule number density per cm³ and \( T \) is temperature in degree Kelvin.

111) CARRUTHERS (1962)

\[ \gamma = \mu \left\{ 1.11 \times 10^{-7} n(S_2) + 7 \times 10^{-9} n(O_2) \right\} \]
where \( u \) is electron energy in \( eV \), \( n(N_2) \) and \( n(O_2) \)
are molecular number densities of \( N_2 \) and \( O_2 \) gases per \( cm^3 \).

iv) a) HILL and BOWHILL (1977)

\[
\gamma_{en} = 1.6 \times 10^5 \text{ p (dynes/cm}^2) \text{ assuming } T = 220^o\text{K}
\]

b) PACK and PHILPES (1961)

\[
\gamma_{en} = 6.6 \times 10^4 \text{ p in the D-region where } p \text{ is in dynes/cm}^2
\]

v) DAVIES (1965)

\[
\gamma = 6 \times 10^4 \exp \left[ - \frac{(h-100)}{6.7} \right] \text{ h in km}
\]

vi) COWLING (1945) and NICOLET (1959)

\[
\nu = 30 N T_e^{-3/2}
\]

where \( \nu \) is the electron ion collision frequency, \( N \)
the electron density per \( cm^3 \) and \( T_e \) the electron
temperature in degrees Kelvin.

vii) BANKS (1965, 1966)

\[
\nu = 3.62 N T_e^{-3/2} \ln \left[ \frac{1.24 \times 10^4}{N^3} \left( \frac{T_i^3 + T_e}{T_i + T_e} \right)^{1/2} \right]
\]

\( T_i \) is ion temperature in degree Kelvin. This expression
was revised as \( \nu = 5.04 N T_e^{-3/2} \).
viii) SAHA & VENKATAGIRI (1975) considered the effect of
the earth's magnetic field and unequal \( T_e \) and \( T_i \) in
arriving at a suitable empirical relation for electron-
ion collision frequency from their work on cosmic radio
noise dispersive absorption. It is \( \nu_{ei} = 150 \times T_e^{-3/2} \).

ix) FRIEDRICH & TORKAR (1983) found the proportionality
constant 'K' between collision frequency and pressure
from the D-region collision frequencies derived from
rocket flights. For the monoenergetic electron they
found that

\[
\nu = K \cdot p = (2.63 \times 10^{-6} T_e^{-0.5} + 4.94 \times 10^{-5}) \times p \text{ s}^{-1}
\]

where \( p \) in \( \text{Nt/m}^2 \). They found the variation of \( K \) with
temperature in the range \( 7.10 \times 10^5 \text{Nt}^{-1} \text{m}^2 \text{s}^{-1} \) and \( 6.47 \times 10^5 \text{Nt}^{-1} \text{m}^2 \text{s}^{-1} \) for \( 150^\circ \) and \( 300^\circ K \). From the weighted mean
of all rocket data, they found

\[
K = 6.24 \pm 0.41 \times 10^5 \text{Nt}^{-1} \text{m}^2 \text{s}^{-1} \text{ at mean}
temperature } T = 213^\circ K. \text{ SMITH et al (1978) and BENNETT}
et al (1972) found the value of \( K \) equal to \( 6.3 \times 10^5 \)
and \( 9.9 \times 10^5 \text{Nt}^{-1} \text{m}^2 \text{s}^{-1} \) respectively, MECKLEY (1974)
showed that \( K \) increases with height.

According to BANKS (1966),

\[
\nu_{eo} = [0] (1.88 \times 10^{-10} T_e^4) (1 \pm 0.3)
\]

\( \nu_{ei} \) and \( \nu_{eo} \) are not important in the lower ionosphere.
Fig. 1.9 gives $\gamma_m$ at different latitudes in different seasons (THAHE et al. 1968).

Fig. 1.9: Height variation of collision frequency adopted for (a) $60^0N$ and $69^0N$ summer and autumn (b) $60^0S$ and $69^0S$ winter and spring (c) low and mid-latitude all year (after THAHE et al. 1968).

In our work of $N(h)$ profile modelling from measured $I_{dB}$ data, we have found satisfactory

$$\gamma_{\text{eff}} = 6.5 \times 10^5 \ p \ \text{N/m}^2$$

where $p$ is atmospheric pressure given by CIRA 1972 model. It is essentially equal to $\gamma_m$ since $\gamma_e$ and $\gamma_o$ are not important in the lower ionosphere as compared to $\gamma_m$ (DATTA et al. 1983; GUPTA and KADIA 1979).
1.12 DIURNAL VARIATION OF ABSORPTION

Assuming the Chapman-layer electron density distribution given by

\[ N = N_0 \exp \left[ \frac{1}{a} - \sec \lambda \exp(-\delta) \right] \quad \ldots \quad \ldots \quad (1.41) \]

where \( a = (h - h_m)/H_m \), \( h_m \) is the height of maximum electron production, \( N_0 \) is the maximum electron density for the overhead sun, and \( H \) is scale height, APPLETON (1937) gave the expression for total nondeviative absorption suffered by the wave in one complete vertical transit as

\[ L = \frac{4\pi N_0 H V}{mc} \sqrt{2\pi e^{\exp(1)}} \frac{\cos^{3/2} \lambda}{(\omega \pm \omega_L)^{3/2}} \quad \ldots \quad \ldots \quad (1.42) \]

for quasi-longitudinal propagation, where \( V \) is electron collision frequency at the level of maximum electron production. According to this equation, the nondeviative absorption should vary with the time of the day as \( 3/2 \) power of \( \cos \lambda \) under ideal conditions of Chapman-layer (1931). The experimental results of total absorption show that the exponent of \( \cos \lambda \) varies from 0.5 to 1.5 depending on the local ionospheric conditions of a place. For a fixed frequency, one can write

\[ L = L_0 \cos^n \lambda \quad \ldots \quad \ldots \quad (1.43) \]

where \( L_0 \) is absorption at \( \lambda = 0 \) and it includes all the
constant quantities as in eq. (1.43).

The exponent $n$ in eq. (1.43) is obtained by finding the slope of the line given by the equation

$$\log L = \log L_0 + n \log \cos \chi \quad \cdots \cdots \ (1.44)$$

From the intercept on the ordinate axis of $\log L$, one can find the value of the subsolar absorption $L_0$. An alternative expression developed at CRPL (now NOAA, USA) to fit the observations near sunrise and sunset is

$$L = A + B \cos \chi \quad \cdots \cdots \ (1.45)$$

The term $A$ justifies the absorption resulting from the ionisation due to photodetachment processes (VAN ZANDT & KNECHT 1964). The factor $B$ represents sensitivity of absorption to the solar zenith distance and it falls off at high latitudes (DAVIES 1960). $A$ may be taken as constant nighttime absorption, and not the mathematical value obtained by extrapolation of the linear fit.

1.13 seasonal variation

According to the $\cos \chi$ -law of variation, the monthly mean noon absorption normalized to a certain solar activity should be maximum in midsummer and minimum in midwinter. However, at midlatitudes it is found that in winter the absorption is higher than expected from the above consideration. On some days in a winter month, the absorption becomes abnormally
high, resulting in a high value of monthly mean. It is also true when anomalous enhancement of absorption on some days occur in one longitude zone, there are anomalous reductions of absorption on the same days in another longitude zone (THOMAS 1962), both at mid-latitudes, although there is on the average a mid-latitude 'Winter anomaly' of enhanced absorption. This amounts to saying that anomalous enhancements of absorption do occur frequently at all mid-latitudes, but with a lag of some days from east to west. In radio communication, the use of high frequencies for increased foF2 in winter more than compensates for the higher absorption in winter at mid-latitudes. References of this winter anomaly are made to the works of CHAPMAN & BAKER 1965, BENSON & JONES (1965), DIRMINGER et al (1966), SCHMIDT (1966), LAUTER & TAUBENHEIM (1970), KOTADIA & PATEL (1971) and Patel et al (1973). It has now been fairly well established that the anomalous increase of absorption in winter at mid-latitudes is due to the increase in D-layer ionisation as a result of change in gas composition and the geopotential height brought about in some way by the wind circulation system in the mesosphere and the stratosphere (PATEL 1975). This phenomenon has attracted special attention of the scientists working on the International Middle Atmosphere Program (IMA 1982-1985).

In contrast to the 'Winter anomaly' in absorption at mid-latitudes, there is a 'Summer anomaly' at low latitudes near
the equator, but this summer anomaly pertains to reduced absorption, compared to that normally expected (SHIRKE and HENRY 1967, GHANALINGAN 1969, RAO and RAO 1972, GUPTA and KOTADIA 1976).

1.14  SOLAR CYCLE VARIATIONS

The ionospheric absorption exhibits a marked 11-year solar cycle variation. It increases by a factor of about 1.5 from sunspot minimum to sunspot maximum. The yearly running mean of \( L \) shows a linear relation with \( R_e \) of the form

\[
L = a (1 + b R_e)
\]

The best experimental value of \( b \) for nondeviative absorption is found to be about 0.004 at midlatitudes. There can be, however, scatter in the plot of monthly value of \( L \) and \( R_e \) which may be due to seasonal irregularities. The value of \( b \) for different months may vary from 0.01 to 0.0045. The most extensive series of absorption data in existence is that (since 1935) for the Radio Research Station (now Rutherford and Appleton Laboratory) at Slough, England, U.K. APPLETON & PIGGOTT (1954) found that the value of \( b \) was highest in summer (\( b = 13.2 \times 10^{-3} \)) and lowest in February (\( b = 7.0 \times 10^{-3} \)) for absorption at noon on 4.0 MHz. The corresponding values of \( a \) varied from 1.0 naper in November-December to 1.5 napers.
in July-August. Results for other low and midlatitude are given in Chapter III. It is interesting to note here that the solar cycle variation of maximum electron density in the E-region follows a different pattern from that of absorption with regard to the sensitivity to solar activity, the value of $b$ for $N_{eq}$ variation being almost uniform over all the months. GHANALINGAM (1974) gave a relation for the variation of absorption with solar X-ray flux $U_x$ as

$$L(2.2 \text{ MHz}, \chi = 10^6) = 4.3 + (11.2 \pm 0.5) \sqrt{U_x}$$

for $1-8$ A° $U_x$ in the range $0.1 < U_x < 2.0 \times 10^{-3} \text{ ergs/cm}^2/$sec.

PARAMESWARAN and KRISHNAHOTHI (1978) found

$$L_{dB} = 42.2 + 6.4 \sqrt{U_x} \text{ for } U_x (1-8 \text{ A°}) < 10^{-3} \text{ ergs/cm}^2/$sec.

RASER (1952) gave an approximate formula, based on absorption measurements at Slough, for nondeviative oblique incidence absorption at midlatitudes as

$$L_{dB} = 4.30(1+0.0035 \text{ R}_x^2) \cos^{0.75} \chi \sec i_p \frac{1}{(f \pm f_L)^2} \quad \cdots (1.47)$$

deibels per hop, where $i_p$ is angle of incidence at the D-layer and $f$ is the oblique-incidence wave-frequency in MHz. The above formula does not take into account the winter anomaly.

For predicting statistical performance index in h.f.
communication via the ionosphere, LUCAS and HICK (1966) gave the following expressions:

Daytime:

\[
I_{dB} = \frac{677 \times 2n \sec 1 (1 + 0.0037 R_s) \left[ \cos(0.881 \chi) \right]^{3/4}}{(f + f_R)^{1.96} + 10.2} \quad \ldots \quad (1.48)
\]

Nighttime:

\[
I_{dB} = \frac{677 \times 2n \sec 1 \left[ 0.025 \left( 1 + 0.013 R_s \right) \right]}{(f + f_R)^{1.96} + 10.2} \quad \ldots \quad (1.49)
\]

where \( n \) = no. of hops, \( \chi \) = angle of incidence at ionosphere, \( R_s \) = yearly running average sunspot number, \( \chi \) = solar zenith angle, \( f \) = operating frequency, \( f_R \) = gyrofrequency at height of reflection. In the above, constant nighttime absorption (free from \( \chi \)) is assumed.

The above three equations serve as a good guide for average behaviour. However, due allowance has to be made to accommodate latitudinal and seasonal anomalies of absorption.

1.15 FREQUENCY-DEPENDENCE OF ABSORPTION

The nondeviative absorption of ordinary wave measured on different frequencies at the same solar zenith distance should, according to the eqs. (1.33) and (1.34), vary inversely as the square of the effective frequency \((f \pm f_R)\) when \( \gamma < \omega \). This can be expressed as
where $A$ is a constant, depending on $\int n \gamma dh$ in the D-layer. Near the critical frequency of a layer and where $\gamma$ and $\omega$ are comparable (DAVISS and HAGG 1955), the inverse-frequency squared law does not hold good. The increase in absorption near the critical frequency of the E and F1 layers due to the deviative absorption was demonstrated by APPLETON and FINGSTT (1954) by simultaneous measurement of absorption at different frequencies (see Fig. 1, 10). To account for the deviative and nondeviative absorption, BIELE ET AL (1962) suggested the expression:

$$L = \frac{A}{(f + f_L)^2} + B \phi \left( \frac{f}{f_L}, \frac{f_x}{f_y} \right) \quad \ldots \quad (1.51)$$

in which the first term gives the Q-wave nondeviative absorption and the second term gives the deviative absorption. $\phi \left( \frac{f_x}{f_L}, \frac{f_x}{f_y} \right)$, a function tabulated by the authors, determines the "proximity effect." $f_y$ is the cusp frequency of the C type sporadic-S. It is possible to calculate the constant $B$ from the values lying outside the line obtained by plotting $1/L$ against $(f + f_L)$ and from the tabulated function $\phi$. The straight line enables to find the constant $A$ as well as the gyrofrequency $f_L$. It is shown by several workers that
the exponent $n$ in the cosine-law of diurnal variation of non-deviative absorption or the $A$-figure is practically the same for different frequencies, which fact implies that the non-deviative absorption mostly occurs in the D-region, although the reflection of the waves may occur at different heights in the E and F-regions. This distinction is also inferred from the solar cycle variation of $A_{mK}$ and $L$. ELLISON and DAVIES (1955) had earlier given the empirical formula for absorption as

$$L = A \left( \frac{\sin \phi}{f - f_0} \right)^n + B \left( f - f_0 \right)^2 \times 7 \quad \ldots \quad (1.52)$$

GEORGE (1971) gave the formula for the dependence of noon absorption using measured data at 27 stations throughout the world. According to George, absorption factor is defined as :
\[ A(f) = L(f) \cdot (f + f_L)^2 \text{dB (MHz)}^2 \quad \ldots \ldots (1.53) \]

and the results at different frequencies of the radio waves reflected from the E-layer were approximately related to the A-factor for the internationally accepted standard frequency 2.2 MHz by:

\[ \frac{1}{f_0^E} \cdot \frac{A(f)}{A(2.2)} = \varphi \left( \frac{f}{f_0^E} \right) \quad \ldots \ldots (1.54) \]

where the function \( \varphi \left( \frac{f}{f_0^E} \right) \) is same for all stations and months including winter anomaly zones. He found \( \varphi(f/f_0^E) = 0.3 \) for \( f > f_0^E \) frequencies of waves reflected from the F-layer using phase-integral calculations for a mean model ionosphere which is consistent with the E-layer results, and called \( A(f) = A_T \) in such cases.

SUNDE & BRADLEY (1975) extended George's analysis to other times of the day and showed that the relationships for diurnal and solar cycle variations of an absorption parameter proportional to \( \int N' \text{dH} \) are simpler than those found for \( L(f) \). The function \( \varphi(f/f_0^E) \) is obtained by taking \( f \to \infty \) in equation (1.54) and defining \( A_T \) as the limiting value of \( A(f) \). Eliminating \( A(2.2) \) then gives

\[ \varphi \left( \frac{f}{f_0^E} \right) = \frac{A(f)}{A_T} \quad \ldots \ldots (1.55) \]
where $A_T$ is directly proportional to $\int N \, dh$, the height integrated product of electron density and collision frequency over the whole of the absorbing region. Their plot of $A_T$ vs. $\cos \alpha$ on log scale showed almost same value of the index $n$ for different frequencies. fermi and Bennett (1979) expressed doubt about the validity of George's A-figure at low latitudes, but recently Ferguson (1984) refuted this criticism and confirming that the function involved in George's $A(f)$ was invariant with respect to latitude.

1.16 LATITUDINAL DISTRIBUTION

It is fairly well established that the ionisation in the lower ionosphere is symmetrically distributed in the two hemispheres around the equator. George (1971) studied worldwide distribution of $I(f + f_d)^2 = A$ and found that the absorption A-figure of the quantity $\int N \, dh$ showed a maximum at about 20° dip and a decline near the equator. From an oceanic expedition observations, Shanks and Henry (1967) noted what they called equatorial anomaly of reduction in absorption. This result of geomagnetic control seems rather curious because the main factors affecting the absorption, namely, the electron density and the collision frequency in the D-layer are not known to exhibit geomagnetic control. There are also reports of a semiannual component of variation in absorption near the
equator somewhat similar to that of $f_0 F_2$ (Gunalingham 1969).

1.17 METHODS OF ABSORPTION MEASUREMENTS

(1) Use of vertical incidence pulse technique or Al method

In this method, the amplitudes of radio pulse echoes received after reflection from the ionosphere are recorded. The daytime amplitude is compared with the nighttime amplitude for calculation of absorption. The nighttime absorption below the F-region is assumed to be nearly zero. The corresponding heights of reflection are also compared to allow for the spatial loss which is inversely proportional to distance travelled. Then the absorption is given by

$$L = 20 \log \frac{E_N h^N}{E_D h^D} \text{ (db)}$$

where $E$ denotes amplitude, $h^N$ virtual height of reflection and the subscripts $D$ and $N$ stand for daytime and nighttime. Reference level $E_N h^N$ is standardized from observations during several nights when clear echoes free from spread-F or Be are observed. An excellent description of this method has been given by Piggott et al. (1956), Piggott and Brown (1963). Further details of our measurements by Al method are given in Chapter II.
Use of Cosmic Radio Noise or A2 method

This method was first introduced by MITRA and SHAH (1953) and later improved by LITTLE and LEIBNACH (1959). Galactic radio noise (a widely spread and diffuse source) at frequency above $f_0P_2$ is received using a narrow vertical-beam broadside antenna array. Detailed description of this method is available in the IQSY instruction manual and URI manual. Absorption in the ionosphere is given by

$$L = 10 \log \left( \frac{P_0}{P} \right) \text{ dB} \quad \cdots \cdots (1.57)$$

where $P_0$ is reference power of the noise which would have been received in absence of the ionosphere. The standard quiet-day curve (QDC) for sidereal time is required for reference level and this is usually achieved by measuring radio noise when $f_0P_2$ is minimum in the early morning just before the sunrise effect starts in the F2-ionization, say at 03:00 hr - 05:00 hr LMT. The frequency on which the CRN is recorded will be quite high compared to $f_0P_2$ at this time and there is practically no absorption of CRN in the ionosphere. A mean straight line showing the constant value during the night hours in the plot of absorption at a particular sidereal time against solar hours also gives the QDC of reference power (SARADA and MITRA 1962).

The major source of error in A2 method is in obtaining
a reliable QDC, for which at least a year’s data free of
spread-F are needed. The noise power is usually expressed in
terms of equivalent temperature as

\[ P = kT(BW) \]  \hspace{1cm} (1.58)

where BW is the bandwidth of the noise spectrum, k
Boltzmann constant and T equivalent temperature of noise
source in degree kelvin. The magnitude of the received power
is obtained with the help of a calibration noise-source
apparatus. According to Hud (1966), ionospheric absorption

\[ L_{db} = 10 \log \left( \frac{T_g}{aT_a} \right) \]  \hspace{1cm} (1.59)

where \( a \) is transmission factor (\(< 1\)) in the ionosphere and
\( T_g \) is equivalent temperature of galactic noise source.
Absorption measured by a practical system is

\[ L_{db} = 10 \log \left( \frac{bcoT_g + b(1-c)T_a + (1-b)T_c}{abcT_g + bc(1-a)T_a + b(1-c)T_a + (1-b)T_c} \right) \]  \hspace{1cm} (1.60)

where \( T_a, T_a \) and \( T_c \) represent equivalent temperature for
losses in the ionosphere, antenna and cable respectively and
\( a, b, c \) are the corresponding transmission or efficiency
factors as the case may be. (\( a \) and \( b < 1 \)). At frequencies below
30 MHz, \( T_g \gg T_a, T_c \) and \( T_a \) in which case eq. (1.60) approxi-
mates to eq. (1.59). At higher frequencies, \( T_g \) becomes comparable
to other temperatures and in the case of PCAs and SCNAS when \( a \) becomes very small, the complete expression (1.60) should be used.

The absorption at 30 MHz is very small (few tenths of a db) except in cases of PCAs, so for good accuracy and to study the variation, a frequency as low as practicable should be used depending on the maximum value of \( f_{e2} \) and the transmitter frequency of local or distant broadcasting and communication stations.

(3) Use of oblique incidence CW signals or A3 method

The measurement of field strength of one-hop CW signals reflected from the E-layer provides a simple method for determining absorption in the lower ionosphere. If the transmitter is working for good span of time, day and night, this method can be used to find diurnal, seasonal, day-to-day and solar cycle variations of absorption. If it works for some hours only, these variations at constant solar zenith angle can be studied. It helps in identifying disturbance periods and anomalous absorption. The absorption at oblique incidence is found by comparing the daytime field-strength with the quite nighttime field-strength free from fading and \( E_2 \) scatter. The method is being used in East Germany, West Germany, Czechoslovakia, Netherlands and Japan with considerable success. The frequency of the signal and the distance between the transmitter and the
receiver have to be chosen in such a way that the nighttime signal by one-hop E-layer propagation is not very weak (say not less than 1 db). F-propagation is avoided and multiple echoes are strongly absorbed. Further, the groundwave limits the maximum ionospheric absorption that can be measured, therefore the power of the transmitter must be such that the ground wave is completely attenuated at the distance of one-hop E received signal or antenna system should be so arranged that the ground wave is eliminated. Again, since the ordinary wave suffers less absorption than the extraordinary wave, the antenna should be so designed and oriented that it selects only the ordinary wave. Good surveys of this method and the results achieved by it are given by SCHWEITZER (1966) and LAUTER (1966).

(4) **Use of \( f_{\text{min}} \) as a measure of absorption**

\( f_{\text{min}} \) represents the minimum frequency recorded in the ionogram for which the echo intensity falls below the minimum recording level. It depends on the following factors:

1. The frequency response characteristics of the recording system,
2. The ionospheric absorption and its frequency dependence, and
3. The noise level present,
Using absorption data obtained by Al method at a place where \( f_{\text{min}} \) is also recorded, though slightly differing in time (by about 5 min), a useful relation between \( f_{\text{min}} \) and \( I_{\text{dB}} \) can be established. \( f_{\text{min}} \) can then serve as a reliable parameter in interpolating the missing value of absorption measured by the other method. It is also potentially useful in distinguishing between days of high and low absorption or solar activity, in studying day-to-day variability and seasonal anomaly, SID phenomena and the occurrence of FCAs.

A serious limitation on the use of \( f_{\text{min}} \) as an index of absorption is imposed by the noise level and the interference by the strong MW signals from the broadcasting station. \( f_{\text{min}} \) would be generally reliable if the monthly median values are low, when gain-settings are made in the ionosonde according to the prevailing ionospheric conditions during the day and the night, care should be taken that a particular gain-setting is done at the same time every day for meaningful interpretation of \( f_{\text{min}} \) data. Reservations on the use of \( f_{\text{min}} \) as an index of ionospheric absorption were pointed out by KOSAI and GUPTA (1976) from a study of \( f_{\text{min}} \) and absorption over a half solar cycle. They found better agreement of \( f_{\text{min}} \) with absorption on the lower frequency than on the higher frequency, but there are quite a few disagreements also and differences in the relative changes of \( f_{\text{min}} \) as compared to those in absorption.
PIGOTT et al. (1956) and THOMAS (1962) have made a
comparative study of $f_{\text{min}}$ and absorption for midlatitude
European stations and such a study is required to be made
at high and low latitudes. Using $f_{\text{min}}$ data, HARRISON (1960)
reported on the increase of ionization in the D-layer by a
factor \((f_{\text{min}} + f_L)^2 / (f_{\text{min}} + f_L)^2\) that occurred during
the solar flare of 10 July 1959, where $f_{\text{min}}$ is the observed
value during the flare and $f_{\text{min}}$ is the mean of values on
control days during the corresponding time interval. $f_{\text{min}}$
observed by continuous operation of ionosonde during a solar
flare or a solar eclipse helps to trace the whole course of
start, growth, decay and recovery phase of the associated
ionospheric phenomena. $f_{\text{min}}$ data are also useful in detecting
PCAs and black-outs. SINGH and HIGASHIMURA (1969), GREGORY
and MANSON (1970) and SHRESTHA (1971) studied winter anomaly
in absorption using $f_{\text{min}}$ data.

(5) Relative merits and demerits of the different methods

The $A1$ method of measurement of absorption is probably
the most reliable and is recommended as the primary system.
It is possible to measure nondeviatiative absorption in the D-
layer and deviative absorption in the E-layer by selecting
the frequency of the exploring radio wave below the critical
frequency of the E-layer, such that the reflection occurs in
the E-layer. Even at frequency $> f_{\text{ce}}$, the nondeviative absorption
mostly occurs in the D-layer. At frequencies generally employed in the A1 method, the total absorption is greater than 10 db. Multifrequency absorption measurements by A1 method are very useful in studying the structure of the lower ionosphere.

The A2 method (by BIOMETER) is generally recommended for absorption work at high latitudes. The correlation between absorption measured by A1 and A2 methods is not good, probably because of the unpredictable deviative absorption of GN occurring in the F2-layer (TAKEMIWA et al 1966). This method has been in use in low latitudes also (Ahmedabad and Delhi) for quite some time on frequency 21-25 MHz. At frequencies greater than 30 MHz, the absorption is very small (< 1 db) which is comparable to errors involved in the determination of absorption and so this A2 method is generally not of much value at v.h.f. The method has been effectively used in solar flare patrolling.

The A3 method is perhaps the simplest one. The correlation between the absorption measured by A1 and A3 method is used at oblique incidence in the LF and MF band over distances of less than 500 km for one-hop S-reflection, but beyond the ground-wave cut-off. The A1 and A3 methods have been successfully used in the study of day-to-day variations, seasonal anomalies and solar cycle variations of absorption
but not so widely for the study of the effects of solar flares or energetic proton events. The A1 and A3 methods are sensitive to small changes in absorption but they cannot easily be used to measure large changes in it. MITRA (1970) has given a good review of the different techniques of absorption measurements and their scientific implications. The three methods were summarised critically by Prof. Rawer (1966) in his concluding remarks on a symposium on ionospheric absorption. The UAG Report 57 (NOAA, U.S.A.), a 'Manual on Ionospheric Absorption Measurements' edited by Prof. Rawer (1976) gives a detailed account of various theories, methods of measurements including Partial Reflection and rocket & satellite-borne techniques, and some important results. Rich material is presented in it by specialists in their respective fields.

1.18 ABSORPTION MEASUREMENTS AT AHMEDABAD

Much of the work on ionospheric absorption has been carried out at midlatitudes and extensive data are collected particularly at European stations, viz. Slough, De Bilt, Dourbes, Juliusruh, Freiburg and Lindau. Some work is done in India, Nigeria, Singapore and Sri Lanka (Ceylon) at low latitudes, but as continuous series hourly observations is maintained over a sizeable period, say at least one solar cycle. In recognition of the key position of Ahmedabad (23° N, 72.6° E, Mag Dip 3° N) in the ionosphere map of the world and in view
of the fact that not much information is available on the lower ionosphere near tropics, and ambitious programmes were launched for systematic long-term studies of the ionospheric absorption by Al method. The Indian National Committee for URSI have also stressed the need of such work over an extensive period. Regular observations commenced in April 1972 on three frequencies, viz., 1.6, 2.2 and 2.5 MHz, and this work formed a part of the Worldwide Middle Atmosphere Program (MAP) also. The description of the equipment, antenna system, the types of records etc. are given in the next chapter.
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