CHAPTER I
INTRODUCTION TO THE EARTH'S UPPER ATMOSPHERE - IONOSPHERE

1.1. Introduction
The Earth's atmosphere is divided into various regions on the basis of temperature and composition. The atmosphere is composed mainly of molecular Nitrogen (N$_2$), and Oxygen (O$_2$) with smaller amount of water vapour and various trace gases. The behaviour of the earth's atmosphere is governed by the sun as the absorption of the solar radiation is responsible for nearly everything that goes on in the earth's atmosphere. The absorption of solar radiation by various atmospheric constituents determines the temperature structure of the atmosphere [Figure 1.1]. The lowest layer of atmosphere is the troposphere (0-15 km) where the temperature falls off throughout the region until it reaches a height in the range of 10-17 km known as tropopause. Above the tropopause lies the region called stratosphere. The main constituent in this region is ozone although its contribution is only few millionths of the total ground level pressure. It absorbs solar ultraviolet (UV) radiation, which heats up the stratosphere, and thus the earth's surface is shielded from the otherwise lethal radiation. The region terminates at 45-55 km height, known as stratopause and then the temperature again decreases throughout the region known as mesosphere, which terminates at about 80-85 km known as mesopause. Heat flows towards this level by conduction from above and is removed by radiation in the infrared and visible airglow, and by downward eddy transport into the mesosphere. Mesopause is, in fact, known as the coldest region of the earth's atmosphere. The temperature minimum here is due to the lack of any strong heating mechanism. Ozone cannot exist at this height region as it is rapidly destroyed by photochemical reaction. Since the region from the earth's ground level to say 100 km is homogeneous because of the thorough mixing by convection and turbulence, this region is called homosphere. Whereas, the region above 100 km is said to be heterosphere because of the significantly varying composition. Also, the region above mesopause is known as thermosphere (90-500 km) where shorter wavelength UV radiations are absorbed which are responsible for the high temperature existing...
Figure 1.1. Thermal structure of the earth’s atmosphere.
there. Here, the temperature increases upward as most of the heat liberated is removed by the downward conduction. Finally, the conduction becomes so good that the region of the upper thermosphere is maintained in nearly isothermal condition at a relatively high temperature (1000-2000°K). Above thermosphere, there lies the exosphere where the ionized particles are constrained by the earth's magnetic field. Finally, there is the magnetosphere, the region in which the earth's magnetic field controls the dynamics of the atmosphere.

1.2. Ionosphere

The upper atmosphere contains free electrons and ions produced by ionizing radiations from the Sun and from the Earth's space environment. The ionized part of the earth's upper atmosphere that contains significant number of free electrons and positive ions is termed as ionosphere. Above 60 km altitude, electrons are sufficiently dense to influence the propagation of radio waves, giving the ionosphere much of its practical importance. The name ionosphere was coined by R. Watson-Watt in 1926, and came into common use about 1932. Since then it has been a subject for many investigators. The ionosphere is formed by the ionization of atmospheric gases such as N₂, O₂, and O. It extends down to perhaps 50 km and thus overlaps the ozonosphere; the symbols D, E, F1 and F2 are used to distinguish the various regions of the ionosphere. It has no well-defined upper boundary, but merges into (or may be extended to include) the heliosphere, where neutral and ionized helium are important constituents. Within any region, distinct "layers" or "ledges" of ionization (as shown in Figure 1.2) may be observed (such as F1, F2 layer in F region) and there may exist a C layer at 50-60 km below the principal D-region ionization. These regions with approximate height and daytime electron densities are given below:

D region, 60-90 km: \( \sim 10^8-10^{10} m^{-3} \) (\( 10^2-10^4 \) cm⁻³)

E region, 90-140 km: \( \sim 10^{11} m^{-3} \) (\( 10^6 \) cm⁻³)

F1 region, 140-180 km: \( \sim 10^{12}-10^{13} m^{-3} \) (\( 10^7-10^6 \) cm⁻³)

F2 region, maximum variable around 300 km: upto \( \sim 10^{12} m^{-3} \) (\( 10^6 \) cm⁻³)

The ionization in the D-region is produced through X-rays (< 10Å), Lyman \( \alpha \)
Figure 1.2. It shows the various regions of the earth's ionosphere and its electron concentration.
(1216 Å), EUV (< 1118 Å) and galactic cosmic rays. The primary ions produced by the above process are NO⁺, O₂⁺, N₂⁺, O₂⁻. The D-region is a region of weakly ionized plasma with a large number-density of neutral species and complex ion-interchange and electron attachment and detachment reactions. The importance of the latter processes (electron attachment and detachment) is the most distinguishing feature of the D-region. The D-region can be considered as the lowest part of the ionosphere where water vapor, cluster ions, both positive and negative, tend to dominate the composition and chemistry.

The normal E-region is produced by the sun's X-rays and ultraviolet radiations. During the daytime, production of charged particles is balanced by the dissociative recombination that act as a sink for charged particles. The E-layer extends over the altitude range 90-140 km. It is in this region where strong electric currents are generated by the dynamo process which greatly affects the physical processes in the E-region. Typical nighttime and daytime electron density profiles, produced as a result of the absorption of solar EUV radiation is shown in figure 1.3.

The partially ionized portion of the upper atmosphere is called the ionosphere and the rate of change of electron concentration (N) in the ionosphere is given by the equation of "continuity" or "balance", whose terms represent the effects of various processes which alter the electron concentration N. Thus, within a cell of unit volume if the transport processes result in a net drift velocity V, then the change due to transport is the divergence of the flux NV and the continuity equation can be written as

$$\frac{\delta N}{\delta t} = q - L(N) - \text{div}(NV)$$

where, q and L represent production and loss rate respectively. When the rate of electron loss is determined by the dissociative recombination reaction, the ions are mainly molecular and the square-law loss formula (αN²) applies whereas in the case when ion-atom interchange reaction controls the rate of loss, the linear formula (βN) applies and ions are mostly atomic in nature.

The two dominant ions in the 90-140 km height range are O₂⁺ and NO⁺. Other primary species are N₂, O₂, O. Although molecular nitrogen is abundant in
Figure 1.3. Normal electron density distributions at extremes of the solar cycle (Johnson, 1965).
the atmosphere there is a scarcity of \( \text{N}_2^+ \) ions in all regions of the ionosphere since they react very quickly with oxygen. The predominant mechanism responsible for the loss of ionization in this part of lower ionosphere is due to the dissociative recombination:

\[
\text{NO}^+ + e \rightarrow \text{N} + \text{O}
\]

\[
\text{O}_2^+ + e \rightarrow \text{O} + \text{O}
\]

The nature of the ionosphere will depend on the rate at which these reactions take place. If \( N \) is the electron density and \( N_{in} \) is the number of ionized molecules then the loss rate of electrons in the E-region is given by

\[
L = \alpha \cdot N \cdot N_{in}
\]

where \( \alpha \) is called the recombination coefficient. Typical value of \( \alpha \) is \( 5 \times 10^{-11} \text{m}^3 \text{s}^{-1} \) throughout the E-region.

The most probable explanation for the existence of night time E-region peak is that the long lived metallic ions, such as Fe\(^+\), Mg\(^+\), Na\(^+\) etc., dominate the night time recombination processes once most shorter lived NO\(^+\) and O\(^+\) ions have disappeared. In situ measurements (Prakash et al., 1970) have shown the presence of ionized layer in the form of NO\(^+\).

The primary ion in the F region is O\(^+\) which is converted into molecular ion by charge exchange reaction:

\[
\text{O}^+ + \text{O}_2 \rightarrow \text{O}_2^+ + \text{O}
\]

\[
\text{O}^+ + \text{N}_2 \rightarrow \text{NO}^+ + \text{N}
\]

The dynamics of the ionosphere over this region is mostly explained by assuming that there is no transport of ionization. Thus, equation (1.1) reduces to

\[
\frac{dN}{dt} = q - l(N)
\]
Which relates the change in electron concentration \( N \) to the ionization production rate \( q \) and the loss rate \( I \). The loss rate in the F-region is given by

\[
I_\ell = \beta N
\]  

Where \( \beta \) is the attachment coefficient.

The F2 peak can be observed in the number density profile of ions and electrons at an altitude of \( \sim 250 \text{ km} \) and above, where the chemical loss rate of ions becomes comparable to the diffusion rate. F2 region is the upward extension of F1 region. In the F2 region, the radiation of wavelength \( \lambda < 90 \text{ nm} \) is mainly responsible for ionization. The F2 layer survives in the night whereas D and F1 layer vanishes by night. Also, the electron density of E-region becomes very weak by night.

1.3. Equatorial ionosphere:
The equatorial ionosphere is unique in the sense that it provides a variety of fascinating physical phenomena: Equatorial Electrojet (EEJ), occurrence of Equatorial Spread F (ESF), Equatorial Sporadic E \( (E_s) \), blanketing sporadic E \( (E\alpha) \), Midnight Temperature Maximum (MTM), and the development of Equatorial Ionization Anomaly (EIA). The dynamics of the F-region is affected by both electric field and neutral winds. In addition, diffusion of ionization along the geomagnetic field \( (B) \) lines also plays a major role in the ionization distribution. Over the equator, the geomagnetic field lines are horizontal and the vertical plasma motion at F-region heights is purely due to electrodynamic drift as the field \( (magnetic) \) aligned plasma flow has no component in the vertical direction. In the F-region, when electron density is taken as a function of latitude, a pronounced 'trough' centered on the magnetic dip equator with crests at \( \sim 15^\circ \) to \( 20^\circ \) north and south is observed \( [\text{Appleton}, 1946] \) and this phenomena is known as equatorial ionization anomaly \( (EIA) \) or Appleton anomaly. This anomaly exists during most of the day and is most pronounced around sunset but disappears after midnight \( [\text{Rastogi}, 1959] \). The causative mechanism for this phenomena is the electrodynamic vertical drift \( \mathbf{E} \times \mathbf{B} \) \( (E \) is the electric field and \( B \) is the geomagnetic induction) of F-region plasma followed by plasma diffusion along the magnetic field lines to form the so called equatorial fountain \( [\text{Mitra}, 1946; \text{Martyn}, 1947; \text{Duncan}, 1960; \text{Coley et al.}, 1990] \) (see Figure...
Figure 1.4. Electrodynamic lifting and fountain effect in the equatorial F region. The upward drift over the magnetic equator is due to an eastward electric field communicated from the F region. The arrows show the flux of ionization in direction and magnitude. The computation is for noon conditions.
1.4). In the low-latitude ionosphere, magnetic field lines are assumed to be equipotential because of the large symmetry in parallel and perpendicular conductivities. Global zonal electric field (E) produced due to the E-region dynamo in conjunction with the earth's horizontal magnetic field creates a vertical polarization electric field (E_p) at the equatorial E region (since the Hall mobility of electrons, at E-region heights, is much greater than that of ions for an applied electric field). This polarization field can drive large amounts of current through Hall drift of electrons in the east-west direction. The resulting intense current is Equatorial electrojet (EEJ).

Another interesting phenomena of equatorial ionosphere is the occurrence of spread F in the F-region. Spread F is a nighttime phenomena which can be seen in ionograms as diffused echoes generated by various plasma instability processes, of which Rayleigh-Taylor instability plays the dominant role in triggering the equatorial spread F. It is discussed in detail in a later section. Spread F can be classified mainly into three categories [McNicol et al., 1956] on the basis of its appearance in the ionograms (Figure 1.5). They are:

1. Range spread F- When the diffuseness is principally along the horizontal part of the F-region trace giving rise to the ambiguity in height but critical frequencies are identifiable
2. Frequency spread F-When the spreading is maximum at frequencies close to the penetration frequency causing ambiguity regarding the identification of critical frequency of F2 layer (f_p,F2) while the trace is sharp and clear at lower frequencies.
3. Complete spread F- Occasionally, spreading is seen to be equally prominent in the height scale in the entire frequency range of the observed ionogram.

The source of the eastward electric field (E_e), near the equator, is believed to be tidal motions at mid-latitudes, which also produce the S_n current system. Horizontal neutral winds, at E-region heights, are capable of driving ions across magnetic field lines through collisions, thereby establishing polarization electric field [Matsushita, 1973; Matsushita, 1977]. It has been shown that field-aligned currents linking the northern and southern hemispheres are important in equalizing the electric fields at conjugate points. The global electric field generated due to the
Figure 1.5. Typical ionograms showing (a) Normal F-layer trace, (b) Range spread F, (c) Frequency spread F and, (d) Complete spread F.
action of tidal wind plays an important role in transporting the ions across the field lines in the ionosphere (ionospheric dynamo). Stening (1977) has shown that the intense east-west current found over the magnetic equator (the equatorial electrojet) is driven predominantly by dynamo action in the 30-60° latitude region. Investigators have examined the morphology of the E-region zonal electric field [Woodman, 1970; Fejer, 1981; Coley and Heelis, 1989; Fejer et al., 1991, 1995; Maynard et al., 1995]. The zonal electric field is generally eastward during the day and westward at night, causing the equatorial ionosphere to rise and fall, respectively. Strong electric fields and plasma density gradients, produced in part by atmospheric gravity waves, introduce instabilities that give rise to plasma irregularities on a wide range of scales. Day-to-day variability of the winds and electric fields, associated with variability in atmospheric tides, planetary waves, and gravity waves, causes variability in the ionosphere and in ionospheric irregularities.

The equatorial ionosphere is also coupled with high-latitude regions and with the plasmasphere and magnetosphere. Magnetospheric electric fields and currents that couple into the high-latitude ionosphere penetrate to the equator, especially during magnetic storms (discussed more in chapter VII) and other disturbances, where they affect the ionospheric electron density, plasma irregularities, and the equatorial electrojet current.

1.4. Geomagnetic $S_q$ variations
The regular diurnal variation in the earth's magnetic field was first noted by Graham [1724]. Later, Balfour Stewart [1882] proposed that they are caused due to the electric currents flowing in the upper atmosphere which is generated as a result of air motion. Air motion caused by atmospheric tides sweeps ions and electrons across the earth's magnetic field lines. This induces electric fields, which in turn produce the current flow. Further investigations of the ionosphere was made by Appleton and Barnett (1925), and Breit and Tuve (1928) which led to the development of dynamo theory which was developed principally by Chapman and Bartels (1940). Continuous observation of the geomagnetic field at any location will show comparatively smooth and regular variations on some days and irregular variations on some other day. The days of first kind are considered to be magnetically quiet or calm whereas the days of second kind are called magnetically active or disturbed.
The extreme case of magnetic disturbances is known as magnetic storm. The degree of magnetic disturbance varies from day to day over a wide range and also from place to place, with the disturbance field superposed on the regular daily magnetic variations.

The worldwide \( S_q \) current system in the ionospheric dynamo region is driven by electromotive forces generated by wind motions on a global scale. The wind will carry ions along with it leaving behind the electrons whose collision frequency is very much less than its gyrofrequency. This wind induced ion motion will lead to a charge separation resulting in electric field and current, whose values varies with latitude and longitude. Thus, a global system of electric field is produced leading to a divergence free global scale current system. Over the equator, the electrostatic field is eastward in daytime and westward during nighttime. \( S_q \) current system depends on the electrical conductivity of the ionosphere (which varies diurnally), as well as on the tidal winds. Tidal wind causes atmospheric motion relative to the geomagnetic field lines. Under geomagnetically quiet conditions these winds and their associated currents can be separated into two component, one due to solar heating (\( S_{q_s} \), quiet day solar) and other due to the gravitational force of the moon (\( L_{q_l} \), quiet day lunar) variations. The amplitude of \( L_{q_l} \) is one tenth of the \( S_q \) and the value of \( S_q \) can vary significantly from day to day, with season and solar activity. Apart from the quiet day daily variation (\( S_q \)) of geomagnetic field, there is an 'additional disturbance daily variation SD' during disturbed or stormy periods. Additional disturbance daily variations 'SD' can be obtained by deducting \( S_q \) from \( S_{\delta_d} \) (actual daily variation on disturbed days). \( S_q \) variations is greatest during the daytime having a sharp peak shortly before noon whereas \( S_q \) variations is mainly diurnal with an early morning peak.

1.5. Dynamo theory
The ionospheric E-region consists of weakly ionized, collisional, magnetized, multi-species plasma, whose electrodynamics is strongly influenced by the physical characteristics of the surrounding medium. The dynamics of the whole ionosphere is subjected to changes frequently by the solar and meteorological phenomena. The periodic heating and cooling of the atmosphere by the solar radiations produces
daily atmospheric oscillations, which propagate upwards into the ionosphere, inducing tidal winds. The current system generated by the tidal motion is called \( S_\text{q} \), meaning the variation related to the solar day under quiet geomagnetic conditions.

According to the dynamo theory, tidal winds produce the requisite motion of the charged particles (electrons and ions) in the ionosphere and induces an electric field \( \mathbf{E} = -\left( \nabla \times \mathbf{B} \right) \) by dynamo action, thereby causing electric current flow which is responsible for geomagnetic \( S_\text{q} \) variations. Here, \( \mathbf{V} \) is the wind velocity, \( \mathbf{B} \) is the geomagnetic induction field vector and \( \mathbf{E} \) is perpendicular to both \( \mathbf{V} \) and \( \mathbf{B} \). Using the generalized form of Ohm's law, the current density \( j \) as a function of electric field \( E \) can be written as

\[
j = \sigma \cdot E \quad (1.5a)
\]

where \( \sigma \) is the tensor conductivity. It is a tensor because of the anisotropy introduced by the earth's magnetic field. The divergence free condition for current flow is given by the equation

\[
\nabla \cdot \mathbf{j} = 0 \quad (1.5b)
\]

The current produced by the tidal winds will not satisfy the current continuity in the ionospheric layer. At any point where \( \nabla \cdot \mathbf{j} \neq 0 \), electric charges will accumulate and an electrostatic-polarisation field \( -\nabla \phi \) (\( \phi \) denoting the electric potential) is set up by the space charge effect (electrons moving separately from ions) so that the total field \( E = -\nabla \phi + \left( \mathbf{V} \times \mathbf{B} \right) \) adjusts itself until the current flow is non-divergent.

Over the dip equator, \( \mathbf{V} \times \mathbf{B} \) is very negligible because \( \mathbf{B} \) is nearly horizontal and the current is driven almost entirely by the field of the space charge and hence it is supposed that the conducting layer is a thin insulated slab having two dimensions only and bounded above or below by horizontal planes and that only the horizontal components of the fields are responsible for driving the horizontal currents.

### 1.6. Electrical conductivity of the Ionosphere

As mentioned earlier, the diurnal variation of the geomagnetic fields are caused by currents in the ionosphere. These currents are produced by the motion of charged
particles in the ionized medium and thus the magnitude of these currents depends on the electrical conductivity of the ionosphere which is a weakly ionized plasma.

In order to understand the dynamics of the E-region it is necessary to know the altitude variation of the electrical conductivities of the different species. The conductivity is anisotropic because of the geomagnetic field. Electrical conductivity is the ratio between current density $j$ (in A/m) and electric field $E$ (in V/m):
\[
\sigma = \frac{i}{E} = \frac{Nev}{E},
\]
where $v$ is the velocity of the relevant charged particles, $N$ their concentration and $e$ the charge on each particle.

In the absence of a magnetic field, or if the electric field is along the magnetic field (B), $v = F/mv = Ee/mv$. Thus,
\[
\sigma = \frac{Ne^2}{mv}.
\]
where,
- $m$ = mass of each charge particle
- $v$ = mean collision frequency of each particle with neutral particle and with each other

Hence, the conductivity due to ion and electron currents is given by
\[
\sigma_0 = \left( \frac{N_e}{m_e v_e} + \frac{N_i}{m_i v_i} \right) e^2
\]
(1.6a)
where
- $m_i$ = mean ionic mass
- $m_e$ = mass of electron
- $N_i$ = total ion concentration
- $N_e$ = total electron concentration

This is called the direct or longitudinal conductivity. At greater heights (above the F2 peak) the 'direct' conductivity $\sigma_0$ given by equation (1.6a) becomes very large, as collision frequency of electron ($v_e$) and the collision frequency of ion ($v_i$) become very small. The conductivity then depends on ion-electron collisions instead of ion-neutral collisions.
Figure 1.6a. Electron density $N_e$, Parallel, Pedersen and Hall conductivities $\sigma_0$, $\sigma_1$, and $\sigma_2$, with $B_0 = 2.93 \times 10^5$ Tesla, $F10.7 = 140 \times 10^{-22}$ W m$^{-2}$ Hz$^{-1}$, and $\chi = 0^\circ$ (Richmond, 1973).

For electric fields (or electric field components) orthogonal to the magnetic field (B) we define two conductivities to deal with currents flowing respectively parallel and perpendicular to the electric field. These are represented by the symbols $\sigma_1$ and $\sigma_2$, called the Pedersen (transverse) and Hall conductivities respectively and is given by

$$\sigma_1 = \left[ \frac{N_e}{m_e v_e} \left( \frac{v_e^2}{v_e^2 + \omega_e^2} \right) + \frac{N_i}{m_i v_i} \left( \frac{v_i^2}{v_i^2 + \omega_i^2} \right) \right] e^2$$  \hspace{1cm} (1.6b)

$$\sigma_2 = \left[ \frac{N_e}{m_e v_e} \left( \frac{\omega_e v_e}{v_e^2 + \omega_e^2} \right) - \frac{N_i}{m_i v_i} \left( \frac{\omega_i v_i}{v_i^2 + \omega_i^2} \right) \right] e^2$$  \hspace{1cm} (1.6c)

In the above expressions $\omega_e$ and $\omega_i$ are the electron and ion gyrofrequencies and $v_e$
and \( v \), are the respective collision frequencies. The electron charge, \( e \), is considered as positive. Here, the Pedersen and Hall conductivities peak at E region heights whereas the direct conductivity continues to increase with altitude (see figure 1.6a). If the electric field is expressed in terms of a vector quantity then,

\[ E = iE_x + jE_y + kE_z \]

Where \( E_x, E_y \) and \( E_z \) are components of electric field and \( i, j \) and \( k \) are unit vectors, the \( z \)-axis being along the magnetic field (as shown in figure 1.6b), then the current vector is given by

\[ j = ij_x + jj_x + kj_x = \begin{pmatrix} \sigma_1 & \sigma_2 & 0 \\ -\sigma_2 & \sigma_1 & 0 \\ 0 & 0 & \sigma_0 \end{pmatrix} \begin{pmatrix} E_x \\ E_y \\ E_z \end{pmatrix} \]  

(1.6d)

Over the magnetic equator, the magnetic field lines are horizontal which means that the medium is bounded in the \( y \) direction (as shown in figure 1.6c). Thus from equation 1.6b, 1.6c and 1.6d,

\[ \begin{align*}
    j_x &= \sigma_1 E_x + \sigma_2 E_y \\
    j_y &= -\sigma_2 E_x + \sigma_1 E_y
\end{align*} \]  

(1.6e)

Under equilibrium conditions, the Pedersen current of ions has to be balanced by the Hall current of electrons generated by \( E_x \) to maintain charge neutrality of plasma. Hence, \( E_x \sigma_1 = E_x \sigma_2 \), and, by substitution,

\[ j_x = \left( \sigma_1 + \frac{\sigma_2^2}{\sigma_1} \right) E_x = \sigma_3 E_x \]

where \( \sigma_3 \) is the Cowling conductivity. Thus a large current flows over the equator due to the extremely large value of \( \sigma_3 \), which is comparable to \( \sigma_n \) and this large current is known as "equatorial electrojet" which is confined to the region a few degrees in width.

The simple model of the dynamo region is therefore a relatively thin, horizontally stratified layer.

1.6.1. Layer Conductivities in the ionosphere

_Baker and Martyn_ (1953) first mentioned about the height integrated "layer"
Figure 1.6b. Charged particle motions in a magnetic field.

Figure 1.6c. Currents and fields in a horizontal slab of ionization at the equator.
conductivities in the actual ionosphere. These arise because of the limited vertical extent of the conducting layer in the ionospheric E region. If an electric field is generated in the ionosphere by dynamo action, the resulting current $\sigma \cdot E$ may contain a vertical component. If so, charges will accumulate at the boundaries of the conducting layer because the current cannot flow into the region of low conductivity. These "polarization" charges will modify the electric field $E$ until the resulting flow is horizontal. Thus, eliminating the vertical electric field under the assumption of zero vertical current, the three dimensional tensor $\sigma$ can be replaced by a two dimensional tensor $\sigma'$, representing the "layer conductivity" whose components depend on the magnetic dip angle $I$. Using co-ordinates $x$, $y$ for the magnetic southward and eastward directions, one can write the layer conductivity as

$$\sigma' = \begin{bmatrix} \sigma_{x'x'} & \sigma_{x'y'} \\ -\sigma_{y'x'} & \sigma_{y'y'} \end{bmatrix}$$

where,

$$\sigma_{x'x'} = \frac{\sigma_0 \sigma_1}{\sigma_0 \sin^2 I + \sigma_1 \cos^2 I} \approx \frac{\sigma_1}{\sin^2 I} \quad (1.7)$$

$$\sigma_{y'y'} = \frac{\sigma_0 \sigma_2 \sin I}{\sigma_0 \sin^2 I + \sigma_1 \cos^2 I} \approx \frac{\sigma_2}{\sin I} \quad (1.8)$$

$$\sigma_{y'x'} = \frac{\sigma_2 \cos^2 I}{\sigma_0 \sin^2 I + \sigma_1 \cos^2 I} + \sigma_1 \approx \sigma_1 \quad (1.9)$$

The approximations given at the right arise because usually $\sigma_0 \gg \sigma_1$ or $\sigma_2$, but near the magnetic equator where $I = 0$, they are not valid and we have

$$\sigma_{x'x'} = \sigma_0 \quad (1.10)$$

$$\sigma_{y'y'} = 0 \quad (1.11)$$

$$\sigma_{y'x'} = \frac{\sigma_1^2 + \sigma_2^2}{\sigma_1} = \sigma_3 \quad (1.12)$$

Two physical consequences follow from these latter results.
First, the very high conductivity $\sigma_{xx} = \sigma_0$ along the field lines ensures that these lines are approximately electric equipotentials. Second, the east-west conductivity $\sigma_{yy}$ at the magnetic equator is very large, being the cowling conductivity $\sigma_2$ which is comparable to $\sigma_0$. This highly conducting strip along the magnetic equator carries a large current known as the "Equatorial Electrojet", which is confined to the region a few degrees in width, where

$$\sigma_0 \sin^2 \theta \ll \sigma_1 \cos^2 \theta$$

The effects of currents as observed at the ground can be calculated most conveniently with the use of "integrated layer conductivities.

$$\Sigma_1 = \int \sigma_1 \, dh$$

and

$$\Sigma_2 = \int \sigma_2 \, dh$$

where, the integration is made vertically through the conducting region.

1.7. Equatorial electrojet (EEJ) and counter electrojet (CEJ)

The existence of narrow band of intense east-west current flowing in the equatorial E region of the ionosphere is termed as Equatorial Electrojet (EEJ). The name 'Equatorial Electrojet' (EEJ) was first suggested by Chapman (1951), much later than its initial discovery in terms of magnetic field signatures. Near the magnetic equator, the perturbations in the horizontal component of the geomagnetic field ($\Delta H$) is directly proportional to the intensity of the overhead east-west current and the perturbations in the vertical component of the geomagnetic field ($\Delta Z$) is proportional to the gradient of the east-west current. Hence, the latitudinal variation of the $\Delta H$ and $\Delta Z$ can be interpreted in terms of a narrow high intensity current centred at ~100 km in altitude with a latitudinal extent of ±3° about the dip equator. The tidal oscillations of the earth's atmosphere in conjunction with earth's magnetic field, generates electric fields (the dynamo fields) in the ionosphere and these fields, in turn, drive electric current through the ionosphere (Hirano, 1952; Baker and
Martyn, 1953, Fejer, 1953]. The first indication of the EEJ current was obtained from ground based magnetometer records of the horizontal component (H) of the earth's field. The strength of the equatorial electrojet current depends on various ionospheric parameters like the eastward electric field strength, the intensity of geomagnetic field, neutral winds, the electron density, ambient temperature, collision frequencies of the charged particles, conductivities, etc. The main features of the EEJ have been presented and discussed in various papers based on ground based and rocket measurements [Singer et al., 1951; Onwumechili, 1959; Cahill, 1959; Maynard and Cahill, 1965; Davis et al., 1967; Sastry, 1968; Prakash et al., 1969, 1970, 1972; Pfaff et al., 1982]. The primary east-west electric field $E_x$ generates a Hall drift of electrons and a Pedersen drift of ions. The differential Hall drift of electrons in the vertical direction tends to set up a vertical polarization electric field ($E_z$) because of the sharp decrease in the Hall mobility ($\mu_z$) above 120 km. This field in turn sets up a vertical Pedersen current (due to ion drift) which is given by $\sigma_z E_z$, where $\sigma_z$ is the Pedersen conductivity. Under equilibrium conditions, this Pedersen current of ions has to equal the Hall current of electrons to maintain the charge neutrality of the plasma. Mathematically, it can be expressed as

$$\sigma_z E_z = \sigma_{zp} E_z \quad \text{i.e.} \quad E_z = \frac{\sigma_{zp}}{\sigma_z} E_z$$

or, $N_q \mu_z E_z = N_q \mu_z E_z$

Where $\sigma_z$ is the Hall conductivity, $\mu_z$ and $\mu_z$ are the Pedersen and Hall mobilities; $N$, the concentration of ions and electrons; and $q$, the charge of electron and ion with appropriate sign.

The electrojet current ($j_z$) in the east-west direction consists of both Pedersen current, $\sigma_z E_z$ and Hall current, $\sigma_{zp} E_z$. Thus,

$$j_z = \sigma_z E_z + \sigma_{zp} E_z$$

Using equation (1.15), we have

$$j_z = \sigma_z E_z + \sigma_z \left( \frac{\sigma_{zp}}{\sigma_z} E_z \right)$$

(1.17a)
\[ j = \sigma \frac{\sigma_2}{\sigma_1} E \quad (1.17b) \]

or \[ j = \sigma \frac{E}{\sigma_1} \quad (1.18) \]

\( \sigma \left( = \sigma + \frac{\sigma_2^2}{\sigma_1} \right) \) is known as Cowling conductivity which maximizes at the dip equator and at a height about 100 km. Since \( \sigma_1 \ll \sigma_2 \) and \( E \ll E_j \) in the E-region (90-140 km), the electrojet current can be considered essentially as the Hall current driven by \( E_j \). Here, \( \sigma_1 \) represents the Pedersen conductivity in the direction of the electric field perpendicular to the magnetic field and \( \sigma_2 \) is the Hall conductivity perpendicular to both electric and magnetic fields.

The electrojet current flows eastward during daytime and westward during nighttime, but the strength of nighttime current is weak as compared to daytime because of the reduction in electron density during night hours. The westward electron drift velocity (corresponding to eastward current) is about 100 metres per second. In order to explain the equatorial electrojet (EEJ) phenomena various theoretical models have been developed among which thin-shell model developed by Baker and Martyn (1953) was the first one which estimates the ionospheric conductivities in the equatorial E region. On the basis of this model various models were further developed by many workers by including the effects of vertical current (Untiedt, 1967; Sugitara and Porras, 1969; Richmond, 1973a) and neutral wind (Richmond, 1973a,b; Forbes and Lindzen, 1976; Gagnepain et al., 1977; Reddy and Devasia, 1981).

One of the most interesting aspects of equatorial electrojet is the reversal of the normal eastward flow of electrojet current to westward flow in the evening hours, on some quiet days. This phenomena is indicated by the depression in the horizontal component of the geomagnetic field, \( H \), below its nighttime level and this feature was termed as Counter electrojet (CEJ) by Gouin and Mayaud (1967) to denote the reversal of currents. The importance of counter electrojet studies was greatly increased when it was shown experimentally that these events are concurrent with the reversal of ionospheric drift direction and the disappearance of \( F_m \) over the magnetic equator [Rastogi et al., 1971]. Also, Bhargava and Sastri (1977) showed
that the critical frequency of F2 layer (f_c,F_2) shows remarkable variations during CEJ and non-CEJ days. Sometimes, depression in the deviation of H without going below the nighttime level is also observed, which is termed as partial-counter electrojet.

1.8. Plasma instabilities in the Equatorial Electrojet region

1.8.1. Equatorial electrojet irregularities

The equatorial E-region irregularities are closely associated with the east-west electric field, which drives the electrojet. The E-region irregularities are anisotropic and scatter radio waves when the incident wave vector becomes nearly perpendicular to the earth’s magnetic field B. These field aligned irregularities drifting in the east-west direction will be used as radar targets through Bragg’s scattering process. Two types of radars are used for irregularity studies (i) Monostatic (pulsed) and (ii) Bistatic (CW and pulsed). The bistatic radars have the transmitter and receiving antenna at different locations whereas monostatic radars use the same antenna for transmitting and receiving the radio wave. For backscatter radar, the angle (θ) between the transmitted and received radio waves is 180° and thus for backscatter observation (θ=180°) the scale size of the scatterers must have a wavelength equal to one half the radar wavelength and must propagate in a direction within the radar beam. Rocket measurements by Prakash et al. (1969, 1970, 1972); Pfaff et al. (1982), as well as radar measurements by Cohen and Bowles (1967) and Balsley (1969) showed that there are two distinct classes of irregularities in the Equatorial electrojet region. The first (now generally referred to as type I or two-stream) has been explained by Farley (1963) and others as being caused by a type of two stream plasma instability or Farley-Buneman instability which sets in whenever the drift velocity of electrons (relative to ions) exceeds the local ion acoustic velocity. The second class (type II) is apparently caused by a gradient drift or \( \vec{E} \times \vec{B} \) instability, which sets in whenever there is an ionization gradient of the right sign and magnitude in the presence of crossed electric and magnetic fields; the gradient should be perpendicular to the magnetic field, but it can be parallel or perpendicular to the electric field [Simon, 1963; Maeda et al., 1963; Reid, 1968; Register and D’Angelo, 1970, Whitehead, 1971].
Some important features of Type I irregularities:

(i) Type I irregularities are characterized by phase velocities which is nearly equal to the ion-acoustic velocity, with slight dependence on wavelength and density gradient, and with a small spread. These waves are generated only when the mean electron drift velocity relative to the ion velocity exceeds the ion-acoustic velocity, but otherwise the wave velocity is independent of the drift velocity.

(ii) The type I echoes as observed by radar appear more or less simultaneously over a range of radar elevation angles, including vertical, once the drift velocity threshold is exceeded.

(iii) The intensity of the type I waves depends strongly on the propagation direction, with those propagating horizontally, parallel to the mean electron flow, being the strongest.

Some important features of Type II irregularities:

(i) Type II irregularities are observed by radar for even very small electron drift velocities whenever the density gradient has the proper sign ($\nabla N \cdot \mathbf{E} > 0$) and they are characterized by a broad Doppler spectrum whose mean value is roughly $\mathbf{k} \cdot \mathbf{V}_d$ where $\mathbf{V}_d$ is the electron drift velocity relative to the ions and $\mathbf{k}$ is $2k_{\text{radar}}$ for backscatter observations (electron drift towards the radar generates waves which produce echoes with a positive Doppler shift).

(ii) The intensity of Type II waves is almost isotropic in the plane perpendicular to the magnetic field. In contrast to type I irregularities, the line of sight velocity of type II irregularities, as measured with the VHF backscatter radar, shows an elevation angle dependence. The phase velocity is approximately proportional to the cosine of the radar elevation angle.

(iii) The strength of the returned scattering signal from the type II irregularities at 50 MHz is proportional to the square of the electron drift velocity, at least for velocities not exceeding the two-stream threshold.

The Doppler spectrum which provides information on the phase velocity and lifetimes of plasma wave scatterers is used as an important diagnostic tool to study
the various instability mechanisms and non linear processes that govern the generation and evolution of E-region irregularities in the ionosphere. Instabilities may arise when the medium becomes perturbed either by chance or by means of some external source. In nature, this external source may be wind in the neutral air or gravity. This perturbation will disturb other aspects of the medium and if one of these consequences has the effect of enhancing the original perturbation then they will rejuvenate and will tend to grow larger. In the equatorial ionospheric E region, the free energy due to strong currents and steep ionization density gradients tend to make equatorial electrojet plasma more unstable.

1.8.2. Two-stream (Farley-Buneman) Instability

The main source of free energy in the equatorial electrojet region comes from the relative motion of electrons with respect to the background ions. Many characteristic features of the type I irregularities have been explained by Farley (1963) using kinetic theory and Buneman (1963) using fluid theory. They showed independently that a collisional, magnetized plasma can be driven unstable by a modified two-stream instability, when streams of electrons and ions differ in velocity by more than the ion-acoustic velocity \( C_s \). These type I waves propagate nearly perpendicular to the magnetic field. Schematic representation of the basic physical mechanism driving the linear Farley-Buneman instability is shown in Figure 1.7. This figure shows a typical day time equatorial electrojet configuration, with an upward directed polarization field and westward electron flow.

In the equatorial electrojet region since the collision frequency of ions and electrons with the neutrals differs, any small perturbation in the medium will tend to induce charge separation. But, in order to maintain the charge neutrality small longitudinal electric fields are set up, thus coupling the density and electric field perturbations. The vectors at the bottom of the Figure 1.7 indicate two forces, ion-inertia and pressure, acting on it. The perturbation can grow if the ion-inertia force dominate over the diffusive pressure force such that more plasma is moved into the region of enhanced density, thus amplifying the initial disturbances.

The detailed theory [Register and D'Angelo, 1970; Sudan et al., 1973; Fejer et al., 1975] shows that the instability applies to waves propagating within a cone of
Figure 1.7. Schematic representation of the basic physical mechanism driving linear Farley-Buneman instability.
angle $\theta$ and is given by

$$V_d \cos(\theta) = C_s (1 + \psi)$$  \hspace{1cm} (1.19)

Where, $V_d$ is the relative drift velocity between electrons and ions ($\vec{V}_d = \vec{V}_e - \vec{V}_i$). $C_s$ is the ion-acoustic velocity, $C_s = K_B \left[ \left( \frac{T_e + T_i}{m_i} \right)^{1/2} \right]$ with $T_e$, $T_i$ the electron and ion temperature, $K_B$ is the Boltzmann's constant and $m_i$ the mean ionic mass and

$$\psi = \frac{V_e V_i}{\omega_e \omega_i} \left( \sin^2 \alpha + \frac{\omega_i^2}{\omega_e^2} \cos^2 \alpha \right)$$  \hspace{1cm} (1.20)

where, $\alpha$ is the angle between the wave and the magnetic field, and the subscripted $\nu$ and $\omega$ are the collision and gyro frequencies for the electrons and the ions. For propagation normal to the field, $\alpha = 90^\circ$ and $\psi$ is $\approx 0.3$ in the E-region, but its value increases rapidly as $\alpha$ moves away from $90^\circ$ because the gyro-frequency of electron ($\omega_e$) is 100 times as large as its collision frequency ($\nu_e$) in the E-region. That's why the waves generated by this mechanism propagate normal to the magnetic field and also the threshold velocity difference is close to the ion-acoustic velocity.

1.8.3. Gradient drift (Rayleigh-Taylor) instability

Type II irregularities are caused by a gradient drift or $E \times B$ instability. The gradient drift mechanism is an example of Rayleigh-Taylor instability (which occurs when a heavier fluid overlies a lighter one). Simon and Hoh (1963) were the first to study this type of irregularities for Laboratory plasma and many features of the type II irregularities are explained further by this instability [Rogister and D’Angelo, 1970; Rogister, 1972; Farley and Balsley, 1973]. The plasma can be unstable provided there is a plasma density gradient of zero order oriented in right direction relative to the electric field driving the electrons. That is the plasma is unstable for an upward electron density gradient and westward moving electrons (typical daytime situation) of sufficient velocity. Reversing the gradient or drift velocity destroys the instability; reversing both (as can be the case during night) restores it.

Figure 1.8 shows the schematic representation of the physical mechanism.
Figure 1.8. Schematic representation of the basic physical mechanism driving linear gradient drift instability in typical daytime equatorial conditions.
driving the linear gradient drift instability in typical daytime equatorial electrojet conditions, with a background density gradient and vertical electric field both directed upwardly. Small perturbation in the ionospheric plasma lead to charge separation producing a small polarization electric field ($\delta E$) which are almost out of phase with the density perturbation (i.e., westward electric field will be associated with density enhancement and eastward electric field with density depletion). Due to the presence of magnetic field, there is now an additional force $\delta E \times B$ such that more plasma drifts into the regions of density enhancement. The results obtained by Balsley and Farley (1971) support the gradient drift instability hypothesis but suggest that the instability generates only large wavelength irregularities, and that subsequent nonlinear processes are responsible for the smaller scale sizes detected in radar measurements. The fluid theory of the gradient drift instability shows that there is no threshold condition for the generation of instability [Farley and Balsley, 1973; Fejer et al., 1975b] and for waves propagating horizontally and perpendicular to the magnetic field with horizontal electron drift and vertical electron density gradient, the condition for instability reduces to

$$\omega_r v_e \Omega_i / v_e k H + \omega_r^2 > k^2 C_s^2$$

(1.21)

where,

$$\omega_r = k V_d (1 + v_e v_i / \Omega_e \Omega_i)^{-1}$$

(1.22)

$$C_s^2 = 2kT/m_i$$

(1.23)

$$H = \left( \frac{1}{N \frac{\delta N}{\delta Z}} \right)^{-1}$$

(1.24)

$v_e$, and $\Omega_i$, are the electron and ion collision and gyro frequencies (all defined to be positive). $V_d$ is the electron drift velocity toward the west, $\omega_r$ is the real part of the wave frequency, $N$ is the electron density, $Z$ is measured vertically upward, $k$ (parallel to $V_d$ and perpendicular to $B$) is the wave number $2 \pi / \lambda$, $C_s$ is the ion-acoustic velocity, $T$ is the temperature of both the electrons and ions.

If the wavelength is sufficiently large, the first term (gradient drift) on the left-hand side of equation (1.21) dominates the second term (two-stream) and we have
The longest wavelength of unstable waves is determined by the dissociative recombination and is of the order of a few kilometres during daytime.

1.9. Equatorial Sporadic E ($E_s$) layer

Ionospheric sporadic E refers to E-region traces on ionogram suggesting unusually high E-region electron densities. The equatorial sporadic E is caused by plasma instabilities arising from the Equatorial electrojet (EEJ). According to the Space Environmental Services Center, "Sporadic E ($E_s$) is transient, localized patches of relatively high electron density in the E-region of the ionosphere which significantly affect radio wave propagation". Sporadic E can occur during daytime or nighttime, and it varies markedly with latitude. $E_s$ can be associated with thunderstorms, meteor showers, solar activity, and geomagnetic activity. In ionograms they appear as dense patches of ionization at a constant altitude of ~100 km and sometimes it appears as a thin layer of enhanced ionization thereby blanketing the overlying F layer and is known as blanketing $E_s$ [Pigott and Rawer, 1961; Matsushita and Reddy, 1967; Reddy and Rao, 1968; Reddy, 1968; Reddy and Devasia, 1973]. Blanketing $E_s$ or $E_{ss}$ appears in the form of thin layers of very enhanced ionization with sharp electron density gradients which provide the basic requirement for the growth of gradient instabilities in the presence of crossed electric and magnetic field. From radio sounding measurements of the ionosphere, it has been known for sometime that there are irregularities of ionization density associated with the EEJ [Matsushita, 1962]. Equatorial $E_s$ irregularities have been studied much more intensively by means of ground based radars like Ionosonde, backscatter methods and also by rocket measurements [Cohen and Bowles, 1967; Bowles et al., 1963; Peterson et al., 1959, Jackson and Seddon, 1958, Prakash et al., 1971]. The occurrence and cause of sporadic E layer varies with latitude. The different processes involved in the formation of $E_s$ layers are the horizontal convergence of ionization due to vertical shears of horizontal neutral winds in the E-region [Whitehead, 1961, Axford and Cunnold, 1966, MacLeod, 1966], gravity waves [Hooke, 1970; Whitehead, 1971] and tidal motions [Chimonas, 1971]. But the net
effects of the gravity waves and tides are the winds. Rocket measurements provided
evidence that these enhanced layers are the ionization convergence-driven layers
aided by the neutral wind system [Smith, 1970]. Castel and Faynot (1970) observed
that the irregularities first appear in the F-region and then move downward
producing weak $E_s$. Figure 1.9. shows different phenomena associated with
equatorial ionospheric E-region.

Figure 1.9. Shows the typical ionograms for (a) Normal daytime E-layer, (b) Sporadic E ($E_s$), and (3) Blanketing $E_s$, over the magnetic equator.

blanketing sporadic E layers (see figure 1.10) is not applicable at the magnetic
equator where the geomagnetic field (B) lines are horizontal and also the wind
generated polarization field prevents the vertical convergence of ionization needed
for the formation of such layers. The formation of sheet-like ionization irregularities
in mid-latitude E-region has been explained in terms of ion convergence due to the
vertical shear of horizontal neutral winds.

Figure 1.10. Idealized illustration of the wind shear mechanism in the E-region
(Whitehead, 1961).
According to this theory, a horizontal neutral wind with a sinusoidal profile in the vertical direction is capable of transporting positive ions vertically by some electrodynamic process. The wind drags with it the positive ions, for which the collision frequency of ions with neutral (ω_i) exceeds the gyro frequency, so that \( v_m > \omega_m \). As a result, the ions experience a Lorentz force \((V \times B)\), and are driven at an angle \( \theta \) to the wind velocity \( V \), so that they accumulate within the shear. The electrons are quite unaffected by the neutral wind, since collision frequency of electron with neutrals is less than their gyro frequency \( (v_{en} < \omega_{en}) \), but are constrained to move along the magnetic field lines. They move in such a way as to neutralize the space charge set up by the ion motion, so that the accumulated layer consists of neutral ionization. Kato (1973) suggested that the partial short circuit of the polarization fields is possible if the neutral wind shears are strong. Reddy and Devasia (1973) showed theoretically that horizontal shears of neutral winds can generate the required convergence of ionization which can lead to the formation of blanketing \( E_s \) over the magnetic equator. Equatorial sporadic \( E \) is much more predictable than at any other latitude, as the appearance and disappearance is very well coorlated with the appearance and disappearance of equatorial electrojet [Rastogi et al., 1971; Rastogi, 1972a,b; 1973a,b; 1975; Krishna Murthy and Sen Gupta, 1972; Chandra and Rastogi, 1974; Sen Gupta and Krishna Murthy, 1975] and is abbreviated as \( E_{S_{eq}} \). The regular \( E_{S_{eq}} \) is caused by cross-field (or gradient) instability at the base of the E-region (90-140 km), where, the vertical Hall polarization field and the vertical electron density gradients are maximum and in the same direction.

### 1.10. Equatorial Spread F

The equatorial spread \( F \) is characterized by irregularities in the plasma (electron and ion) density and electric field distributions perpendicular to the earth's magnetic field. Electron density irregularities often develop in the ionospheric F-region. The diffuseness of the F-region echoes, generally called Spread F, near the magnetic equator was first noticed by Booker and Wells (1938). Lyon et al. (1960) and Singleton (1960) showed the existence of a belt of high occurrence of spread \( F \) between ±20° magnetic latitude. Dengey (1956) first proposed that the ESF could be
initiated on the bottomside of the F-region by Rayleigh-Taylor (R-T) instability. Dagg (1957) suggested that ESF is caused by irregularities produced first in the E-region and then coupled to F-region through highly conducting magnetic field lines. Later Martyn (1959) suggested that ESF is a manifestation of the E x B gradient drift instability. Thus, various theories have been set forth in order to explain the generation mechanism of ESF. The current understanding of the generation mechanism of the ESF is shown in figure 1.11. As indicated in this figure, the ESF generation depends basically on three factors: (i) the linear growth rate for generalized R-T instability process (as box identified by '1'), (ii) flux tube integrated Pedersen conductivity that controls the non-linear development (as identified by box '2'), and (iii) density perturbations to serve as seeding sources (as box identified by '3'). It is now well accepted that the primary mechanism responsible for the generation of ESF is the collisional Rayleigh-Taylor instability followed by a hierarchy of other instabilities [Haerendel, 1973]. After sunset the E-region begins to recombine and the conductivity becomes too weak to short out any irregularities in the F-region. Due to the recombination (in E-region) and the upward plasma drift (due to E x B) in the F-region, the electron density gradient in the bottomside of F-region begins to steepen. When the altitudes of the F-region is high enough or when the background electron density is steep enough to overcome recombination effects, plasma density fluctuations begin to grow on the bottomside via collisional R-T instability mechanism. These irregularities will in turn form plasma density depletion on the bottomside which will then rise nonlinearly by E x B drift due to polarization field through the F region peak and cause spread F. Numerical simulation studies have also shown that in the nonlinear regime, the development of instability would lead to the growth of plasma density depletion in the bottomside and their evolution into the topside [Scannapieco and Ossakow, 1976; Zalesak and Ossakow, 1980; Ossakow, 1981]. In low latitude region the background eastward electric field plays an important role in uplifting the ionosphere to higher altitude, which is known to be important for the onset of ESF [Farley et al., 1970; Tsunoda and White, 1981; Abdu et al., 1983; Vyas and Chandra, 1991]. In the bottom side F region, the gravitational force and density gradient are antiparallel, providing the necessary condition for the R-T instability [Haerendel, 1973]. The spread F and its related phenomena vary with season, local
Figure 1.11. A schematic representation of the electrodynamic processes playing roles in the instability mechanism leading to the generation of equatorial spread F/plasma bubble phenomena. Ux and ΔUx are the thermospheric zonal wind and its longitudinal/local time gradient, respectively; ∇.EEJ is the equatorial electrojet convergence at sunset; Σp and ΔΣp the integrated Pedersen conductivity and its longitudinal gradient at sunset; Δn/n the inverse of the bottom side density gradient scale length; ΔnL/nL inverse of the longitudinal scale length of the electron density; ASSY.EIA asymmetric equatorial ionization anomaly. The three factors that control the ESF generation are indicated as 1, 2 and 3 in their respective boxes (Abdu, 2001)
time, geographical location and with many other geophysical conditions [Shimazaki, 1959; Herman, 1966; Skinner and Kelleher, 1971; Sastri et al., 1979a,b; Rastogi, 1980; Abdu et al., 1981, 1992; Subbarao and Krishna Murthy, 1994]. The basic mechanism of ESF generation is a combination of the gravitational R-T instability that favours high altitudes with low ion-neutral collision frequency and the ExB drift instability due to the ionospheric east-west electric field in conjunction with the horizontal magnetic field that pushes the ionosphere up. Spread F and its related phenomena have been studied extensively using a variety of observational techniques, including bottom side sounding, top side sounding, ground based radar observation (Woodman and LaHoz, 1976; Tsunoda, 1980), satellite in-situ measurements (Aggson et al., 1995), satellite radio beacon waves (Abdu et al., 1992, Aarons, 1993), and optical methods [Biondi and Sipler, 1985].

1.11. F-region dynamo

Balfour Stewart (1882) proposed that the currents could be driven by upper atmospheric winds that could generate electromotive forces in the conducting medium as they move it through the geomagnetic field, in effect acting as an electric dynamo. In thermospheric wind system, the pressure gradient produced by the solar heating acts as the main driving force for the F-region dynamo. During nighttime, the F-region transverse Pedersen conductivity is greatly diminished, so that F-region dynamo action above 200 km becomes relatively more important. The major F-region current system arising from dynamo action by winds depends on the Pedersen conductivity ($\sigma_t$) which, at F-region heights (where $\omega_i \gg v_i$) is given by

$$\sigma_t = N e v_i / \omega_i = N m_i v_i / B$$  \hspace{1cm} (1.26)

Simplified F-region dynamo equation given by Rishbeth (1971a) is as follows

$$j_i = -N e U_i \left( \frac{v_i}{\omega_i} \right) \sin \lambda$$  \hspace{1cm} (1.27)

$$j_N = N e U_i \left( \frac{v_i}{\omega_i} \right) \csc \lambda$$  \hspace{1cm} (1.28)

where, $j$ is the current density, $\lambda$ the magnetic dip angle, $v_i$ and $\omega_i$ the collision and gyro frequencies of ions, $U$ the neutral wind and subscripts $E, N$ refer to the
geomagnetic eastward and northward direction. Under assumption that the thermospheric wind is controlled by ion-drag and therefore neglecting acceleration, coriolis and viscosity, the current density can be expressed as

\[ j = \frac{B \times \nabla p}{B^2 \sin I} \] (1.29)

where \( B \) is the magnetic field and \( \nabla p \) is the horizontal components of the neutral air pressure gradient. The F-region dynamo has a higher internal impedance and is essentially a current generator, delivering a current given by equation (1.29) which shows that the total current flowing across a line joining two points in the F-region is related to the pressure difference between the two places although \( B \) and \( I \) vary with location. The magnitude of current does not vary much with location or time, except near the maxima and minima of pressure. If the underlying E-region is sunlit and sufficiently conducting, it will short-circuit the polarization field and current continuity will be assured by a flow of field-aligned current between the E and F-regions. Whereas during nighttime, when the E-region conductivity is small the short-circuiting effect disappears and a polarization field builds up. This effect was discussed with artificial ion clouds by Haerendel et al. (1967) and neutral F-region irregularities by Waldmann and Da Rosa (1973). In recent studies it has been suggested that the effects of F-region dynamo are important even during nighttime [Crain et al., 1993]. Also, during nighttime when the E-region ionization almost vanishes, the F-region plasma forms a layer with a well defined lower boundary. The F-region, thus, can therefore be approximated to a slab geometry as shown in figure 1.12. The F-region plasma has a constant, finite Pedersen conductivity inside the slab and zero outside, a constant zonal wind \( U \) is also assumed to be prevalent in this slab. Because of the eastward zonal wind \( U \) forcing the plasma across the magnetic field \( \mathbf{B} \), a polarization electric field \( \mathbf{E} \) gets generated, as shown in figure. The electric field \( \mathbf{E} \) thus induced gives rise to a zonal drift motion \( V_\theta \) in the F-region plasma. It is to be noted here that, during nighttime, the induced zonal motion in plasma is in the same direction (i.e., eastward) as that of the neutral wind \( U \). For the vertical field, the charge separation would occur in such a way that the overall current \( j_z \) becomes zero. Thus.
\[ j_z = \sigma_1 E_z + \sigma_1 U_2 B_z = 0 \]  
(1.30)

It implies that \( E_z = -U_2 B_z \). For an ideal F-region dynamo, the above equation can be generalised and written as \( \vec{F} + \vec{U} \times \vec{B} = 0 \).

---

Figure 1.12. The schematic diagram showing the F-region dynamo action (Kelley, 1989).

The physical meaning of the above equation is that the electromagnetic force \( \vec{F} \) on the plasma vanishes and the thermospheric wind blows freely without ion-drag irrespective of the density and the altitude of the plasma inside F-region. Since, during daytime the integrated E-region Pedersen conductivity is comparable or more
than the magnetic field line integrated F-region conductivity, the F-region polarization field can't build and, hence, the F-region ion-drag remains high. The theory of the F-region dynamo is based on the assumption that geomagnetic field lines are equipotentials, and the currents may flow freely along them. Theoretical investigation by Farley (1960) and Reid (1965) support the view that field lines may indeed be regarded as equipotentials for electron distributions of scale size greater than a few kilometer (the scale size being normal to the field line at ionospheric heights). Thus the assumption is valid for large-scale structure (100-1000 km) associated with F-region dynamo, though direct experimental evidence for equipotentiality is difficult to obtain. In F-region, the vertical drift \((E \times B)\) of plasma is very significant over latitudes close to the magnetic equator, whereas at latitudes away from the magnetic equator, the meridional component of the neutral wind contributes significantly to plasma drifts, apart from diffusion.

1.12. Dynamic and Electrodynamic effects

In the ionosphere, atmospheric wave motions couple into plasma dynamics through electrodynamic processes. Atmospheric waves are influenced by winds and wind shears in the atmosphere which are formed as a result of solar heating. The electric fields induced by the ionospheric wind dynamo causes plasma drifts, forming equatorial ionization (Appleton) anomaly in electron density. Electric fields and currents interact strongly with the upper atmosphere [Banks, 1979]. The drifting ions, as they interact collisionally with neutral molecules, exert a force on the air and tend to bring it toward the ion motion. Above ~200 km altitude this effect can be important: at high latitudes, winds are common that approach the rapid velocity of the convecting plasma [Meriwether, 1983], while at low latitudes, where plasma drifts are much smaller, the collisional interaction tends to retard the winds driven by pressure gradients. The heating of upper atmosphere due to currents in the auroral region is also of great importance. The basic principle of equatorial ionospheric electrodynamics is shown in figure 1.13. The heating can make a significant contribution to the upper atmospheric energy budget and can even be the dominant heat source above ~120 km during magnetic storms. As the temperature increases, the upper atmosphere expands, and the drag on near-Earth satellites is
increased, changing their orbits [Joselyn, 1982]. Plasma temperature and chemical reaction rate are also affected by the rapid ion convection through the air. Even at mid and low latitudes plasma drift plays an important role in the ionosphere by raising or lowering the layer into regions of lower or higher neutral density respectively, so that chemical decay is retarded or accelerated.

![Diagram of ionospheric electrodynamics]

Figure 1.13. The basic principles of equatorial ionospheric electrodynamics (Rod Heelis, CEDAR, 2001)

Strong electric field and plasma density gradients, produced by atmospheric gravity waves introduces instabilities that give rise to plasma irregularities on a wide range of scales. The variations in the ionosphere and ionospheric irregularities are mainly caused by the day-to-day variations of the electric field and winds that are associated with the variabilities in atmospheric tides, planetary waves and gravity waves. Electrodynamic effects are different for different longitude sectors of the Earth, associated with longitudinal variations in the latitude of the magnetic dip equator and magnetic field strength as well as with probable longitudinal variations in the
atmospheric wave motions. The equatorial ionosphere is also coupled with higher-latitude regions and with the plasmasphere and magnetosphere. Electrodynamic coupling of the ionosphere with the plasmasphere affects large- and small-scale electric fields. Magnetospheric electric fields and currents that couple into the high-latitude ionosphere penetrate to the equator, especially during magnetic storms and other disturbances, where they affect the ionospheric electron density, plasma irregularities, and the equatorial electrojet current. During magnetic storms the plasma drift effects on the ionosphere are not only intensified but are also supplemented by indirect effects through modification of the neutral atmosphere [Prölls, 1980]. The combination of all these various processes must be responsible for causing both instantaneous and delayed perturbations over low and equatorial thermosphere-ionosphere system [Fejer, 1997]. It is known that even during quiet-time periods, equatorial electric fields and currents often exhibit fluctuations with quasi-periodicities ranging from few minutes to few tens of minutes. These short-term variations can be associated with ionospheric wind dynamo system [Richmond, 1994].

1.13. Present Study

The present study stems on the electrodynamic/neutral dynamical effects observed in the equatorial ionospheric E and F regions by using the data obtained from the ground based experimental techniques like VHF and HF radars at Thumba (near Trivandrum), Ionosondes, located at both Thumba (dip 0.5°N) and SHAR (dip 10°N), and MST radar located at Gadanki (dip 13°N). The study deals with different aspects of the E and F regions of the equatorial ionosphere under various conditions of the magnetic activity, solar activity and seasonal variations. The occurrence of plasma density irregularities during nighttime in the equatorial ionospheric F region is a common phenomena, known as equatorial spread F. To understand this phenomena, an ISTEP (Indian Solar Terrestrial Energy Program) campaign was conducted during the equinoctial months of March-April, 1998 with an aim to study the sustenance, growth and decay of the spread F. Using the $h'$F values for the above period, obtained from the ionograms of Trivandrum and SHAR (Geog. lat. 13.7°N, geog. long. 80.2°E, dip lat. 10°N) an off equatorial station, the thermospheric neutral winds under varying geomagnetic conditions has been derived
by using the method given by *Krishna Murthy et. al.*, 1990. The results so obtained are presented in chapter III, preceded by description of the experimental technique given in chapter II. Chapter III is followed by Chapter IV in which the dynamic and electrodynamic changes brought out by the dusk time solar eclipse of August 11, 1999 in the equatorial ionospheric E and F-regions are highlighted. A unique observational campaign involving ground based ionosonde, VHF and HF radars from the equatorial location of Trivandrum, India was conducted to study the eclipse induced effects in the equatorial ionospheric E and F regions. Chapter V gives the first experimental evidence for the existence of vertical Pedersen current at E-region altitudes. This investigation was carried out on the basis of the data obtained under campaign mode of observation from the Indian MST radar located at Gadanki (dip °13°N), VHF backscatter radar located at Thumba, Trivandrum (dip °0.5°N) and Ionosondes located at both Thumba and Sriharikota, SHAR (dip °10°N) during 17-20 November, 1999. The equatorial E-region shows a variety of phenomena viz. Sporadic E layer, Blanketing E, the dynamics and electrodynamics of which is quite complicated. In order to understand these phenomenon an ISTEP (WG-III) campaign was conducted during June 19-July 07, 2000. The objective of this campaign was to investigate the growth, sustenance and decay of different sporadic E (E,) layers. A case study based on the data obtained from VHF radar and Ionosonde, both located at Trivandrum reveals the role of electron density gradient for the generation of Type II irregularities under different CEJ conditions. The results so obtained are presented in chapter VI. The chapter VII gives the response of equatorial ionosphere in the Indian sector to the major magnetic storm of November 4, 1993. The study presented in this chapter is based on the data sets obtained from the ionosonde and magnetometer networks spanning the region 0.3-34.5°N dip. in the Indian sector. The last chapter VIII includes the discussion and summary of the present work.