Chapter 1

Introduction to Equatorial and Low-latitude Ionosphere-Thermosphere System

1.1 Terrestrial Atmosphere

The terrestrial atmosphere is the gaseous envelope surrounding the surface of earth that sustains life and is unique among the planetary atmospheres known so far. The Earth's atmosphere below the altitude of 15 km is classified as 'lower' atmosphere. between 15-90 km as 'middle' and above 90 km as the 'upper' atmosphere. On similar lines, the atmosphere within ± 30° latitude is termed as 'equatorial and low' latitude, from ± 30° to ± 60° as 'mid' latitude and ± 60° to ± 90° as 'high' latitude atmosphere. However, the exact description of these atmospheric regions is primarily based on their various physical and chemical properties. Based on temperature structure, the atmosphere has been categorized as 'troposphere' (~0-15 km), 'stratosphere' (~15-50 km), 'mesosphere' (50-90 km) and 'thermosphere' (above 90 km). The transition regions separating them are namely the tropopause (~15 km), stratopause (~50 km) and the mesopause (~90 km) respectively. However, these boundaries are not very well defined. The temperature trend in the upper atmosphere, i.e., thermosphere is distinctively different from that in the lower atmosphere. There is a steep temperature gradient and the temperature reaches a maximum at the exobase i.e. ~ 400 km. This temperature, highest in the atmosphere, however is highly variable and depends significantly on the solar activity levels. In the regions above ~400 km the atmosphere becomes relatively less dense and collisionless. therefore nearly isothermal condition prevails there. The thermosphere, though primarily neutral in terms of dominant constituents, exhibits a significant presence of ionized species, which forms the ionosphere.
1.2 The Ionosphere

The ionosphere is that part of the upper atmosphere where free electrons exist in sufficient density to have an appreciable influence on the propagation of radio waves. [Rishbeth and Garriott, 1969; Banks and Kockarts, 1973; Kelley, 1989; Hargreaves, 1992, Rishbeth, 2002]. The presence of ionosphere was first inferred by Marconi through the trans-Atlantic radio wave propagation experiment in 1901. In 1902, Kennelly and Heaviside suggested independently that an ionized layer exists around the earth could guide the electromagnetic waves to great distances. The term ‘ionosphere’ was coined later by Robert Watson-Watt in 1926 to replace its previous name, the ‘Heaviside layer’ (after the physicist Oliver Heaviside). The height of this ionized layer and its properties were measured in a beautiful set of experiments by E. V. Appleton and his associates. For this pioneering work, Appleton received the Nobel Prize in physics in 1947. His early experiments and theoretical work were covered in a lecture delivered in May 1932. It furnished much of our knowledge of the ionosphere until the era of rocket and satellite experiments which began in the 1940s and 1950s. The ionosphere, as understood today is the weakly ionized component of the thermosphere.

The ionosphere is formed mainly due to photo ionization of neutral atoms/molecules by the solar X-ray and ultraviolet (UV) radiation and also by the precipitation of energetic charged particles at higher latitudes. Starlight and cosmic rays also cause minor ionization primarily at low altitudes (~60-90 km). During the day solar X-ray and extreme ultraviolet (EUV) radiation are the main ionization sources. The ionization rate depends on the intensity of the ionizing radiation, atmospheric density, composition, and ionization cross sections of the atmospheric constituents.

The three main atmospheric constituents N\textsubscript{2}, O\textsubscript{2}, and O are ionized by the incoming solar radiation and the primary ions produced thereby are N\textsubscript{2}+, O\textsubscript{2}+, N+, O+, and NO+. A variety of physical, chemical and dynamical processes result in the redistribution of this ionization and leads to the formation of distinct ionization peaks and layers, which are generally denoted by the symbols ‘D’, ‘E’, ‘F\textsubscript{1}’, and ‘F\textsubscript{2}’. The D-region (~60-90 km) corresponds to a sparse layer of polyatomic ion clusters while the E-region (~90-120 km)
corresponds to a moderately dense layer of molecular ions NO+ and O2+. The F-region, on the other hand, corresponds to a dense layer consisting mainly of O+, O2+, NO+, and N2+ ions that sometimes gets split into two distinct F1 and F2 layers during daytime. Regions below and above the F2 peak are called the bottom-side and topside ionosphere respectively. A typical electron density profile showing the various ionospheric layers is shown in figure 1.1

![Fig. 1.1 A typical electron density profile showing different ionospheric layers](image)

The equatorial and low latitude ionosphere often shows the presence of an additional layer, between 500 km and 700 km altitude, now known as the F3 layer. Earlier, this layer was known as the G-layer, which was later named as F3 layer by Balan et al. [1997] since it was not found to involve any new ionization source or neutral species and arises essentially due to the dynamics of the F-layer at low-latitudes. Similarly, another feature often found in the equatorial topside ionosphere is the ionization ledges [Sayers et al., 1963; Sharma and Raghavarao, 1989 and references therein, Uemoto et al., 2006]. The issues concerning the generation of these two characteristic features of the equatorial ionosphere are dealt with in detail in Chapter 6 of the present thesis.
Earth's magnetic field plays a dominant role in the dynamics of electrons and ions in the E and F regions of the ionosphere. In the region below about 90 km, the collision frequencies of electrons and ions are greater than their respective gyro frequencies. Hence, the particle motion is essentially controlled by collisions as if the geomagnetic field were absent. Above ~130 km, on the other hand, the collision frequencies of both electrons and ions are smaller than their respective gyro frequencies and the effect of collisions become relatively unimportant for both the species and the particle motion is mainly determined by the geomagnetic field. In the region between ~90 km and 120 km, electrons are strongly magnetized whereas the ions are strongly influenced by collisions resulting in a differential motion between these two. In response to the neutral dynamics, these latter two regions behave as current (F-region Dynamo) and voltage (E-region Dynamo) generators respectively [Rishbeth, 1971; Heelis et al., 1974], controlling the motion of the plasma therein. The phenomena of E- and F-region dynamo are briefly explained in the following sections.

1.3 F- Region Dynamo

The thermospheric wind system produced by the pressure inequalities in the solar EUV heating is the main driving force for the F region dynamo. Rishbeth [1971] first put forward the idea of F-region having its own dynamo. After sunset, the E-region conductivity becomes too small and hence the F-region plasma forms a layer with a well-defined lower boundary, which can be approximated into slab geometry (Figure 1.2a).

The F-region plasma has a constant, finite Pedersen conductivity inside the slab, and zero outside, and a constant zonal wind is also prevalent in the slab. As a result, vertical polarization electric fields develop at the F-region, giving rise to a zonal drift ($V_B$) of the F-region plasma, which, during nighttime, is in the same direction as that of the neutral wind, i.e., eastward. [Rishbeth, 1971; Heelis et al., 1974]. For the vertical electric field, the charge separation would occur in such a way that the overall current $J_z$ is zero. Hence, we have

$$J_z = \sigma_P E_z + \sigma_p u B = 0$$

which implies
\[ E_z = -u B \quad \ldots \text{1.2} \]

For a perfect F-region dynamo, the above equation can be generalized and written as

\[ E + U \times B = 0 \quad \ldots \text{1.3} \]

This means that during nighttime, the electromagnetic force on the plasma vanishes and the thermospheric wind blows freely without ion-drag. The nighttime observations of zonal plasma drift and neutral wind velocities have actually shown that they are nearly equal. During the day, however, the scenario changes because the integrated E-region conductivity is comparable to or more than the magnetic field-line integrated F-region conductivity. Hence, the F-region field cannot build up and the daytime ion drag remains high. This has significant consequences, the finest example being the *Neutral Anomaly* over low latitudes, which is explained later in this chapter.

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**Fig. 1.2a** Actual and slab geometry of the F-region dynamo. [After Keller, 1989]
During daytime, the magnitude of the electric field is essentially determined by the winds in the E-region and the ionospheric conductivity as explained in the following sections.

1.4 E-Region Dynamo

The E-region dynamo is driven by the tidal oscillations of the atmosphere [Baker and Martyn, 1953; Chapman, 1956]. The term 'tide' in the present context refers to the solar driven wind in this part of the atmosphere. The largest atmospheric tides are the diurnal and semi-diurnal tides. These tides are global in extent and they are generated by the absorption of solar radiation mainly by ozone and water vapor in the lower atmosphere. In the absence of any dissipation or trapping, the kinetic energy per unit volume ($\rho U^2/2$) of the upward propagating tends to remain constant and hence the velocity $U$ increases with altitude. The efficiency of a tidal mode to generate ionospheric currents depends upon its vertical wavelength. A mode of short vertical wavelength tends to produce currents that reverse direction every few kilometers inducing, therefore, very small vertically integrated horizontal currents. Thus only those tidal oscillations which remain in-phase over the ionospheric height range of large conductivity (~90-120 km in the E-region) will produce significant currents in the ionosphere establishing the dynamo process [Baker and Martyn, 1953; Chapman, 1956; Richmond et al., 1976; Rishbeth, 1971, 1997; Rishbeth and Garriot, 1969]. It is well known that the effect of neutral winds together with diurnal and semi-diurnal tidal components in the atmosphere cause currents to flow in the 90 to 120 km altitude region.

Neutral winds with velocity $U$, moving across the geomagnetic field $B$, induce a Lorentz force $q(U \times B)$ on a charge $q$. The extent to which ions and electrons get affected by this force depends on the ratio of their respective collision frequency ($v_{\|}$) to gyro frequency ($\Omega_r$), which in turn is a function of altitude. Above ~90 km only ions drift under the action of wind since $v_{\|} >> \Omega_r$ and $v_{\|} << \Omega_r$. This ion motion tends to accumulate charges at a rate given by $\nabla \cdot (N_q v_{\|})$ where $v_{\|}$ is the ion drift velocity, and hence a polarization electric field $E_{\parallel}$ (\(v_{\parallel}\) represents the zonal direction) is set up. Electrons
move freely along the magnetic field lines to adjust the charge distribution. The vertical electric field $E_p$ thus generated, then causes the ions and electrons to drift across the field lines, such that at steady state the current is divergence free, \textit{i.e.},

$$\nabla \cdot J = \nabla \cdot N_e q (V_i - V_e) = 0 \quad \ldots \quad 1.4$$

where $V_e$ is the electron drift velocity. The current $J$ can be expressed as

$$J = \sigma \cdot E' \quad \ldots \quad 1.5$$

where, $E'$ consists of an induced component $U \times B$ due to the effect of the wind $U$ and a polarization component $E_p$. This polarization field is derived from a potential $\phi$ produced by the charge distribution, \textit{i.e.},

$$E' = E_p + U \times B \quad \ldots \quad 1.6$$

where

$$E_p = -\nabla \phi$$

The potential $\phi$ depends on the global distribution of wind and conductivity.

Since the electric field generated through mechanical motion of the ions/electrons by wind in the presence of geomagnetic field is similar to the so-called ‘Dynamo’ action; this ionospheric process in the E-region is known as the E-region dynamo or the Solar quite ($S_q$) wind dynamo current system. Resulting from this current system is an electrostatic field directed eastward from dawn to dusk at low latitudes. In order to understand why this dynamo process is particularly efficient in the E-region, the variation of the electrical conductivity of the ionosphere has to be understood. A brief account of which is given in the next section. The importance of E-region dynamo is in the fact that it generates the global scale dynamo electric field, which in the presence of the N-S geomagnetic field at the equatorial region gives rise to the Equatorial Electrojet (EEJ) current.

1.5 Electrical Conductivity of Ionosphere and the EEJ

The electrical conductivity of an ionized gas in the presence of a magnetic field ($B$) depends on the number density of the charged particles ($N_e, N_i$), the electrical
charge \( (e) \), mass \( (m) \), the mean collision frequency \( (\nu) \) and the gyro frequency \( (\Omega) \). In the ionosphere, where both electric and magnetic fields are present, the conductivity is determined by the basic equations of motion for electrons and ions under the action of these fields [Cowling, 1945; Baker and Martyn, 1953; Chapman, 1956]. The ionospheric conductivity is anisotropic principally because of the geomagnetic field. In general, the three components of conductivity are defined as (1) longitudinal conductivity \( (\sigma_L) \) for motion parallel to \( B \) and \( E \), (2) Pederson conductivity \( (\sigma_P) \) for motion perpendicular to \( B \) and parallel to \( E \) and (3) Hall conductivity \( (\sigma_H) \) for motion perpendicular to both \( B \) and \( E \).

Theoretical expressions for these conductivities are usually derived from multifluid theory, in which neutrals and the different charged species are treated as separate fluids that interact through collisions and are given as

\[
\sigma_L = Ne^2 \left[ \frac{1}{m_e \nu_e} + \frac{1}{m_i \nu_i} \right]
\]

\[
\sigma_P = Ne^2 \left[ \frac{\nu_e}{m_e (\nu_e^2 + \Omega_e^2)} + \frac{\nu_i}{m_i (\nu_i^2 + \Omega_i^2)} \right]
\]

\[
\sigma_H = Ne^2 \left[ \frac{\Omega_i}{m_i (\Omega_i^2 + \nu_i^2)} - \frac{\Omega_e}{m_e (\Omega_e^2 + \nu_e^2)} \right]
\]

At the equatorial region, where the global scale dynamo electric field \( (E_z) \) is predominantly in the east-west direction and the geomagnetic field in the north-south, the electrons drift vertically with respect to ions. Since the \( E \)-region dynamo is bound by relatively non-conducting regions, a vertical polarization field \( (E_z) \) is set up (Figure 1.2b). This polarization field sets up a Pederson current of ions, so that at equilibrium, the vertical Hall current due to electrons is balanced by the Pederson current due to ions. i.e.,

\[
\sigma_H E_x = \sigma_P E_Z \text{ or } E_z = (\sigma_H / \sigma_P) E_x
\]
Fig. 1.2b The enhancement of Cowling conductivity in slab geometry. [After Kelley, 1989].

The vertical polarization field $E_z$ is, therefore, greater than the primary field by a factor of $(\sigma_H/\sigma_p)$. This induced polarization field $E_z$ drives a Hall current $(\sigma_H E_z)$ in the east-west direction. The total horizontal current in the east-west direction is given by

$$J_z = \sigma_p E_z + \sigma_H E_z$$

or.

$$J_z = [\sigma_p + (\sigma_H^2/\sigma_p)]E_z = \sigma_e E_z$$

$$\sigma_e = 1 + \frac{\sigma_H^2}{\sigma_p}$$

Where, $\sigma_e$ is the Cowling conductivity. Thus, the direct conductivity in the East-West direction is enhanced from $\sigma_p$ to $\sigma_e$ and the current is enhanced by $(\sigma_H^2/\sigma_p)E_z$. This enhanced current that flows in the altitude range 90-120 km at and around $\pm 3^\circ$ of the magnetic equator is known as the *Equatorial Electrojet (EEJ)* [Reddy, 1998 and references therein].

The eastward electric field in the dynamo region has another important consequence at the equatorial region. The eastward electric field at the off-equatorial latitudes is mapped to equatorial F region. After sunrise this zonal electric field causes both electrons and ions at and above $\sim 130$ km over equator to drift upwards. This vertical motion of plasma over dip equator is known as the 'fountain' mechanism wherein the F-region plasma is uplifted only to diffuse along the magnetic field lines aided by the
gravitational and pressure gradient forces. This produces a double-humped latitudinal
distribution of ionization known as the Equatorial Ionization Anomaly (EIA), which is
seen in the peak electron density of the F region foF₂, as well as in the Total Electron
Content (TEC). The strength of the EIA shows one-to-one correlation with the time
integrated strength of the equatorial electrojet confirming that the driving force for both
the EIA and the EEJ is the same [Raghavara et al., 1978], and that is the dynamo
electric field. The EIA, which is primarily a daytime feature, has significant implications
in the generation and evolution of many of the daytime as well as nighttime processes in
the equatorial and low latitude ionosphere. The following sections describe some of these
processes, especially those relevant to the present work.

1.6 Quiet-time Equatorial and Low Latitude Thermosphere-Ionosphere Processes

As is understood today, the upper atmosphere is a two-component system, the
components being the thermosphere (i.e., neutrals) and the ionosphere, (i.e., plasma). Our
comprehension of the processes dominant in this part of the atmosphere can increase
significantly only when the thermosphere and ionosphere are investigated as a coupled
system. As seen in the previous sections, at the core of equatorial ionosphere-
thermosphere system (EITS) is the ionospheric wind dynamo and the associated global
currents and electric fields. The EEJ and the EIA are the two fundamental features of the
equatorial and low-latitude ionosphere, resulting from its unique electrodynamics. Apart
from these, there are important indicators of the internal couplings of EITS, for instance.
the Neutral Anomaly [Hedin and Mayr, 1973] and Equatorial Temperature and Wind
Anomaly (ETWA) [Raghavara et al., 1991; Raghavara et al., 1993]. The ETWA has
very important implications in the thermal budget of the low-latitude thermosphere
ionosphere system, and it basically interlinks the EIA to the Equatorial Spread F (ESF)
phenomena.

EITS is also closely coupled to magnetosphere and lower atmosphere and the
near-Earth space environment as well. As an outcome of this, the EITS exhibits several
unique phenomena occurring with characteristic local time dependencies as shown in
Figure 1.3.
The complex web of interactive processes and major EITs phenomena along with their local time dependences [after Jasiński et al., 2002].

**FIG. 1.3**
An excellent review of the EITS phenomena during quiet and disturbed times is given by Sastri et al. [2002]. It can be seen that the processes, which are a result of the prevailing thermosphere-ionosphere coupling, control almost all dynamical features of the upper atmosphere. Since the present thesis deals mainly with the F-region EITS processes and their inter-relationships, a detailed description of the formation and variability of EIA, ETWA and ESF are given in the next sections.

1.7 The Equatorial Ionization Anomaly

The discovery of EIA was based on the equinoctial data of 1944 (a low sunspot year) when two peaks in electron density were found at ±18° magnetic latitude [Appleton, 1946]. As one would expect the electron density to be maximum over equator and monotonically decreasing on both sides, the observed distribution of ionization was considered to be anomalous, and hence it was called the Equatorial Ionization Anomaly (EIA). The presence of EIA in the low latitude ionosphere was further confirmed for both high and low sunspot years [H. Maeda, 1955; K. Maeda, 1955].

Since then, the EIA had been probed by various techniques like the ground-based ionosondes, satellite borne topside sounders, satellite beacon monitoring of the total electron content (TEC) etc [for example, Croom et al., 1960; Duncan, 1960; Lyons and Thomas, 1963; Rao and Malhotra, 1964; Walker and Ma, 1972; Anderson and Klobuchar 1983]. The EIA was first explained by Mitra [1946], who suggested that ionization produced by the solar UV radiation in the upper atmosphere above the magnetic equator is diffused towards north and south along the magnetic field lines. Later, Martyn [1947] proposed the drift theory and showed that it is actually the vertical drift of plasma over dip equator followed by diffusion along magnetic field lines, which is responsible for the double-humped distribution of electron density. Later it was suggested by Goldberg [1965] that EIA is simply a natural, steady state electron density distribution under the combined influence of gravity, electric and magnetic fields and the production and loss processes. Moffet [1979] reviewed the gross features of EIA taking into consideration the physical and electrodynamical processes that control this phenomenon.
Anderson [1981] combined the production, recombination and the vertical drifts of ionospheric species and brought forth a reasonable realistic model describing the diurnal and latitudinal variation of ionospheric plasma density and hence provided a representation of the EIA development. This study also highlighted the importance of vertical plasma drifts caused by the electric field in the presence of earth’s magnetic field. The zonal electric field \( E \), which is eastward during the day, causes a steady upward plasma drift given by \( E \times B \). During daytime, this zonal electric field is essentially the tide-induced E-region dynamo electric field, which is mapped along the field lines into F region heights. The vertical \( E \times B \) drift is independent of charge and mass. The net result is a fountain of dense equatorial plasma rising under the action of the vertical drift, until the resulting meridional pressure gradient become strong so that the plasma start diffusing down the magnetic field lines assisted by the gravity and poleward winds. This leads to the redistribution of plasma density (Fig. 1.4). The strength of the anomaly thus formed was defined by Rush and Richmond [1973] using a parameter defined as the ratio of electron densities at the crest to that at the trough multiplied by the dip latitude of the crest. Raghavaraao et al. [1978] used this parameter and showed that the development of EIA was related to the strength of the EEJ. They showed a high degree of positive correlation with the integrated electrojet strength and the strength of EIA. This result was quite significant as it directly indicated towards the control of the zonal electric fields on the formation of EIA.

Observations using ground based and topside ionograms and total electron content measurements have shown that there is a large day-to-day variability in the development of EIA. One of the remarkable variability shown by EIA is its latitudinal extent with varying geomagnetic conditions, seasons and solar epochs. Large longitudinal differences are also observed [Garg et al., 1983; Raghavaraao et al., 1988; Sharma and Raghavarao, 1989; Rastogi and Klobuchar, 1990; Abdu et al., 1990; Walker et al., 1994].
In the modeling scenario, as mentioned earlier, a reasonably realistic model for the EIA was first given by Anderson [1981]. Afterwards, numerous experimental and modeling studies have been carried out to understand the various aspects of EIA. Recently, the Sheffield University Plasmasphere Ionosphere model (SUPIM) is used to study the EIA characteristics including its north-south asymmetry [Balan and Bailey, 1995; Balan et al., 1995; 1997]. This model has been successful in quite realistically bringing out the EIA features. This model also revealed the possibility of an additional layer above the F2 peak around the equator, known as the F3 layer, the formation of which is much interlinked to the prevalent vertical \(E\times B\) drift, and hence to the EIA. The SUPIM model simulations presented later in this thesis (Chapter 6) bring out these aspects. Excellent review articles describing the EIA has been presented by Rajaram [1977], Moffet [1979], Anderson [1981], Walker [1981], Abdu et al., [1990], Sastri [1990], Stening [1992], Preble et al., [1994], Bailey et al., [1997], and Rishbeth [2000] on various aspects of the EIA. A detailed description of the general morphology of EIA is given in the subsequent sections.
Height Structure of EIA

The EIA extends throughout the F-region of the ionosphere, starting from about 200 km but with a characteristic altitude dependent manifestation. At lower altitudes below the height of the peak electron density, the EIA crests are shallower and lie farthest from the dip equator. With increase in altitude, the crests gain prominence and move closer to the dip equator such that the crest to trough ratio is maximum around the peak of the F-layer \( (h_mF_2) \). In the topside F-region, the crest amplitudes again decrease and the anomaly is not usually seen above 1000 km [Sastri, 1990]. The locus of the crests at different altitudes follows a geomagnetic field line called the anomaly field line. The EIA also manifests even in the latitudinal distribution of the ionospheric total electron content.

Diurnal Pattern of EIA

The development of EIA exhibits a diurnal pattern that is dependent on the phase of the solar activity cycle. At solar maximum, the formation of the crests takes place around 0900 LT, and the crests continue to develop throughout the day till around 2000 LT, and the crests densities eventually decay later in the night. The significant post-sunset EIA is caused by the complex coupled ion-neutral dynamics near sunset hours, when the F-region dynamo takes control from the E-region dynamo. The extent of the crests tends to stretch farther in latitude during high solar activity phase. During solar minimum, the crests begin to form in the bottomside F region around 0900 LT, as in the case of the solar maximum, but the crests continue to develop only up to ~1600 LT, and disappear by around 2100 LT. The maximum latitudinal extent of the ionization crests is also ~6° less at solar minimum than at the solar maximum epoch [Sastri, 1990]. However, the diurnal development of the anomaly shows pronounced longitudinal differences [Walker, 1981].

Seasonal Pattern of EIA

An established feature of the EIA is the north–south asymmetry of its two crests in their amplitude, crest latitudes, and in the dependence on local time and longitude. At solar maximum as well as solar minimum epochs, the daytime anomaly during solstice
months is asymmetric with a larger crest in the winter hemisphere than in the summer hemisphere. At the solar maximum, the sense of the asymmetry reverses when the anomaly persists well in the night. The asymmetry is manifested in the equinoxes also, though with a marked day-to-day variability. The asymmetry extends well into the topside F-region, where the locus of the anomaly crests in the two hemispheres follow different field lines. The symmetric/asymmetric nature of the EIA crests in a given season is controlled by the F region meridional neutral winds, because of its ability to move the ionospheric plasma along the magnetic field lines depending on its direction (poleward/equatorward) and the height-dependent F-region chemical recombination [Sastri. 1990].

1.8 Equatorial Temperature and Wind Anomaly (ETWA)

The Wind and Temperature Spectrometer (WATS) and the Langmuir probe experiment on board the Dynamic Explorer-2 (DE-2) satellite during high solar activity epoch (1981-1982) provided evidences of anomalous variations of zonal wind and temperature over low latitudes [Raghavaraa et al., 1991; 1993]. The zonal wind was found to exhibit a broad maximum around the dip equator, associated with minima on both sides at ±25° dip latitude. Temperature, on the contrary showed a minimum over dip equator and maxima on both sides. Moreover, the minima of zonal winds and the maxima in the neutral temperature were found to be nearly co-located with the crests of the EIA, while the maximum in zonal wind and the minimum in temperature were found to be co-located with the trough of the EIA. The relative variations characterizing the temperature anomaly in ETWA are between 4-10% [Raghavaraa et al., 1993]. This is quite significant as the extent of diurnal temperature variations over low latitudes itself is only ~13%. The decrease in zonal wind velocities from equatorial maxima to minima on either side had been as large as ~100 m s⁻¹, and the temperature increase amounts to an extent of 50-100K. This was the first experimental evidence where the neutral wind dynamics also was shown to be controlled by the geomagnetic field configuration similar to EIA.

In fact, before this experimental evidence itself, many studies regarding the ion-neutral interactions proposed similar effects in temperature and wind as that of EIA.
while studying the latitudinal distribution of molecular species, i.e., N₂ obtained from OGO-6 data showed a 20% enhancement in the N₂ density at ±17° dip latitude and a decrease over the dip equator at 1700 LT, while N₂ density was found to be maximizing over the dip equator at ~0600 LT. This kind of latitudinal distribution for the neutral species, where the crests and trough were found to be co-located with the EIA crests and trough was considered quite anomalous, and hence this feature was named as the Neutral Anomaly (NA). They tried to explain the NA on the basis of enhanced ion drag offered to the zonal flow of neutrals by the enhanced densities over EIA crests. Hedin and Mayr [1973], through a simple two-dimensional theoretical model dealing with both the viscosity and the ion drag, tried to estimate the diurnal component of temperature and density variations over low latitudes. It was shown that at the crest regions, the energy transport from the hot dayside to the cold nightside by means of the zonal thermospheric flow is damped by the increased ion drag due to enhancement of plasma density at the crests of the EIA. However, this mechanism failed to explain the observed temperature feature in ETWA, because it could only predict a localized warming in the evening sector, contrary to the observations. Fuller-Rowell et al. [1997] did another important study wherein they have theoretically simulated the equatorial thermosphere ionosphere system and tried to explain the observed temperature and wind variations on the basis of chemical heating. Recently, Pant and Sridharan [2001] evaluated the local time variability of ETWA, and tried to quantify the roles of ion drag as well as chemical heating.

Another important aspect related to ETWA is the setting up of a meridional circulation cell. The evidence of such a meridional circulation cell came from observation of significant vertical neutral winds that are upward (downward) at the EIA crests (trough) location [Raghavrao et al., 1993]. This has vital consequences as the vertical winds at the equator often assist the growth of irregularities in the nighttime ionosphere.

The Equatorial ionosphere often becomes turbulent during nighttime, and this unstable condition, leading to irregularities in equatorial F region plasma is referred to as Equatorial Spread-F (ESF). ESF is of much current research interest especially from the
viewpoint of navigation and communication as it can cause severe scintillations even at
the L-band frequencies. The basic instability mechanisms responsible for the generation
of ESF are discussed in the next section.

1.9 Equatorial Spread F (ESF)

ESF is characterized by the presence of a wide spectrum of field-aligned irregularities
extending over nearly seven orders of magnitude in the spatial scale which get generated
under favorable conditions due to a hierarchy of plasma instabilities and manifest in a
variety of forms such as scattered echoes in ionograms, plumes in VHF radar maps, bite-
outs in night OI 630 nm airglow intensity, bubbles in in-situ satellite measurements and
scintillations in VHF/L-band satellite beacon signals [Kelley, 1989; Abdu, 2001 and
references therein]. The collisional Rayleigh-Taylor Instability is the primary
destabilizing mechanism operating in the base of the F-region followed by secondary
instabilities, one feeding on the other. An excellent description of these processes is given
by Kelley [1989].

1.9.1 Linear Theory of the Rayleigh-Taylor Instability

The primary process responsible for the generation of large-scale ESF
irregularities is the Collisional Rayleigh-Taylor (CRT) instability mechanism. After
sunset, in the absence of photoionization, bottomside of the F- region ionosphere
undergoes rapid ion-electron recombination and the electron density decreases resulting
in the steepening of the vertical density gradient. This situation in the post sunset
ionosphere (F- region) is analogous to the case in which lighter fluid supports a heavier
density fluid against gravity. In case of ionosphere, the magnetic field is the ‘light fluid’.

The CRT mechanism can be explained as follows: consider a sinusoidal
perturbation in the ion and electron densities ($n_i$), over the steady state value ($n_0$) in the
F- region along the zonal direction as depicted in Figure 1.5. The $\nabla n_0$ is directed anti-
parallel to the gravity. The magnetic field $B$ is into the plane of the paper. Under the
action of gravitational drift, perturbations in ion and electron densities move eastward
and westward with a gravitational drift velocity of $g/\Omega_i$ and $g/\Omega_e$, respectively:
\( \Omega_i \) and \( \Omega_e \) being the ion and electron gyro frequencies respectively. Since gravitational drift is inversely proportional to the gyro frequency, the perturbation in ion densities moves faster than those of electron and lead to the charge separation. Due to this charge separation, polarization electric fields get generated which are directed eastward in the density depleted and westward in the enhanced regions. These polarization electric fields make the depleted region to drift upward and bring the enhanced density region downward, amplifying the density perturbation. In order to sustain the growth, this process should be faster than the effective lifetime of the ions i.e. \( 1/v_R \) (\( v_R \) is the recombination rate), otherwise the ions would be lost by recombination inhibiting the growth of the perturbation. The plasma-depleted region is called a bubble in analogy to the hydrodynamic case. The upward drift of the plasma ceases at the altitudes where the ambient electron density becomes equal to that inside the bubble. This determines the maximum altitude of plasma bubbles and the altitudinal extent of equatorial spread \( F \).

**Fig. 1.5** (a) Schematic diagram of the plasma analog of the Rayleigh-Taylor instability in the equatorial geometry. (b) Sequential sketches from photos of the hydrodynamic \( R-T \) instability. A lighter fluid initially supports a heavy fluid [After Fejer and Kelley, 1980]
The expression for the linear growth rate $\gamma_e$ for the gravitational R-T instability has been shown [Haerendel, 1973; Ossakow et al., 1979] to be

$$\gamma_e = \frac{1}{L} \frac{g k_z^2}{\nu_m}$$  \hspace{1cm} (1.12)

where $\nu_m$ is the ion-neutral collision frequency, $g$ is the acceleration due to gravity, $L$ is the inverse gradient scale length, given by $L = \left[ \frac{1}{n_0} \left( \frac{dn_0}{dz} \right) \right]^{-1}$. The zonal component of the total wave vector $K$, whose magnitude represents the wave number, is denoted by $k_z$.

The growth rate is shown to be maximum when the perturbation is along the zonal direction and becomes independent of wavelength for horizontally propagating wave [Ossakow, 1979] and hence the eqn. 1.12 becomes

$$\gamma_e = \frac{1}{L} \frac{g}{\nu_m}$$  \hspace{1cm} (1.13)

Apart from gravity, eastward electric field also contributes to the growth of the amplitude of the density perturbation. However, the mechanism is slightly different. Due to the Hall drift ($E \times B / B^2$) the plasma is lifted up to higher altitudes to a region of smaller ion neutral collision frequencies, thereby increasing the growth rate of the R-T instability. This drift is independent of the mass and charge of the species; however a charge separation does occur along the direction of the electric field due to the differences in the Pederson mobility $\left( \frac{1}{eB \Omega} \right)$ of ions and electrons. Here, $\nu_{in}$ can be either ion-neutral or electron-neutral collision frequency $\left( \nu_m, \nu_n \right)$ as the case may be.

Due to this charge separation, eastward polarization electric fields get set up in the density troughs and westward in the density crests. This results in the differential vertical drifts as discussed in the case of gravity, affecting the growth of the perturbation amplitude. The effect of the primary eastward electric field $E_x$ [Ossakow, 1979; Kelley, 1989] in enhancing the growth rate of the R-T instability is given as:
\[ \gamma_e = \frac{1}{L} \frac{E_x}{B} \] \hspace{1cm} \ldots 1.14

In addition to the electric field, neutral parameters like eastward zonal wind \( W_x \) can also drive the instability when the background electron density gradient is westward. The growth rate given as:

\[ \gamma_w = \frac{1}{L} W \left( \frac{v_m}{\Omega_i} \right) \] \hspace{1cm} \ldots 1.15

The vertical wind \( (W_z) \), though small in magnitude, exists in the equatorial ionosphere [Biondi and Sipler, 1985; Raghavarao et al., 1987; Raghavarao et al., 1993; Sekar and Raghavarao 1987]. Thus an expression for the growth rate of the generalized R-T instability including all the above-mentioned parameters is given as [Sekar and Raghavarao, 1987].

\[ \gamma = \frac{1}{L} \left[ g \left( v_m \right) + \frac{E}{B} + W \left( \frac{v_m}{\Omega_i} \right) - W_z \right] \] \hspace{1cm} \ldots 1.16

However, this expression takes care of the linear growth of the R-T instability, which is the primary mechanism for the generation of ESF. A ‘hierarchy of plasma instabilities’ is believed to be the cause for the observed wide range of scale sizes, extending to shorter scale lengths through secondary plasma processes [Haerendel, 1973; Chaturvedi and Kaw, 1976; Costa and Kelley, 1978]. The hierarchy is as follows:

- Collisional R-T instability mechanism driven by the zero order electron density gradient
- The \((E \times B)\) gradient drift instability due to the sharp density gradients set up by the collisional R-T instability mechanism
- Collisionless R-T instability in the region where collisions become negligible and grows due to the sharp density gradients, and finally.
- Kinetic drift waves grow off these irregularities as they attain large amplitudes.
The basic idea of the multi-step (hierarchy) process for the growth of ESF irregularities is now well accepted though there are some differences regarding the details of the processes. By far, the nonlinear theories of ESF have been very successful in explaining different kinds of observations on ESF irregularities at various altitudes from different longitude regions across the globe.

1.9.2 The Background Ionospheric-Thermospheric Conditions Favoring ESF

As is evident from equation 1.16, growth of ESF irregularities is interlinked with the variability of many of the thermospheric and ionospheric parameters, including the zonal and vertical winds. Each of these terms, and their relation with the prevailing background ionospheric and thermospheric conditions is explained in detail in Chapter 5.

During daytime, the electric field is eastward and, therefore plasma drift is upward. The plasma drift reverses to the downward direction at ~1900 LT. However, just before the drift reversal at the time of sunset, the eastward drift is enhanced briefly. This enhancement is ascribed to the post-sunset or the pre-reversal enhancement (PRE) of the eastward electric field. The PRE causes rapid uplifting of the F region, and steepens the bottomside gradient leading to the R-T instability. The vertical uplift of the plasma layer by the eastward electric field helps in increasing the bottomside density gradient.PRE holds the key to the formation of irregularities, and hence an important question to be addressed is what drives the PRE and which parameters control its magnitude. Two main mechanisms have been proposed [Eccles, 1998], as are briefly discussed below.

Mechanism 1

This mechanism was given by Farley et al. [1986] based on a simulation study. In their simulation, they completely suppressed the E-region dynamo although the conductivity and its loading effect on the F-region electrodynamics were retained. They then injected a uniform 200 ms\(^{-1}\) eastward wind. This simulation could produce PRE quite well. Figure 1.6 illustrates the mechanism. The eastward component of the neutral wind in the F region causes a downward electric field (\(\cdot U \times B\)). This electric field is smaller on the dayside because of larger E-region conductivity to which it is connected by the magnetic field. The electric field maps down to the E region and drives a westward
Hall current in the E region. Because of the much-reduced Hall conductivity on the nightside, the Hall current is much larger on the dayside. This results in a collection of polarization charges near the terminator and creates an eastward electric field enhancement on the dayside of the terminator [Farley et al., 1986].

Fig. 1.6 Simplified model of the F-region pre-reversal enhancement of the zonal electric field. The neutral wind blows across the terminator generating a vertical electric field, which is no longer shorted out. This field maps along B to an equatorward electric field in the E layer and tends to drive a westward current on both sides of the terminator. If no current flows in the night-side E region, a negative polarization charge builds up near the terminator, creating the zonal electric field $E_{\phi}$. This $E_{\phi}$ maps back to the F region [after Farley et al., 1986]

Mechanism 2

In this case, the F-region dynamo [Rishbeth, 1981] is considered to be driving the vertical Pedersen current, and the electrojet supplies the current demanded by the F-region dynamo. The rapid drop of conductivity in the equatorial electrojet at sunset coupled with increased current of the F-region dynamo on the night side requires an enhanced zonal electric field or PRE. The PRE draws the electrojet current zonally to meet the current demands of the F-region dynamo on the night side. [Hauerendel et al., 1992]

In the past, there were supporting evidences for both these mechanisms as possible candidates for causing the PRE. Presently, there are indications that it could be due to the mechanism proposed by Rishbeth [Eccles, 1998]. Recently Sridharan et al.,
suggested that the mechanism proposed by Farley et al. [1986] is also equally important for the post sunset lifting of the F-layer. The details of this study are given in Chapter-5. Notwithstanding the mechanism, during the post sunset hours the equatorial F-region gets lifted up due to the enhanced zonal electric field exhibiting significant day-to-day variability. Owing to the smaller collision frequencies at higher altitudes it becomes conducive for the R-T instability to get triggered. The lifting of the F-layer to 300 km and above has been generally construed to be favorable for the triggering of ESF [Kelley and Maruyama, 1992].

Neutral Wind Terms

The role of the zonal wind as a destabilizing factor for the R-T instability is important in the presence of a zonal gradient in electron density (equation 1.16), which is normally the case during sunset. Further, it is the same zonal wind that creates the F region dynamo which in turn is responsible for the pre-reversal enhancement and the subsequent post-sunset height rise of the F-layer. Hysell and Burcham [2002] have shown that the climatological records of the occurrence of ESF with season, solar flux and geomagnetic activity levels and suggested that the climatological behavior of the zonal electric field itself, controlled by the neutral wind behavior, plays a dominant role in the occurrence/ non-occurrence of ESF. Similarly, the in-situ measurements and the theoretical simulations have shown that vertical wind also have a major role in the triggering of ESF [Raghavarao et al., 1987; Sekar and Raghavarao 1987; Sekar et al., 1994]. As mentioned earlier, these vertical winds are often connected with the meridional circulation connected to ETWA, which in turn is connected to EIA. Even though the importance of the ambient ionospheric and thermospheric conditions for the onset of ESF on a given day is well established [Raghavarao et al., 1988; Sridharan et al., 1994; Mendillo et al., 2001; Devasia et al., 2002], the day-to-day variability of the occurrence of ESF remains an enigma. These aspects are discussed in detail in chapter 5.

The Seeding of ESF

Apart from the background ionospheric-thermospheric conditions described above, the occurrence of ESF necessitates the presence of a 'seeding perturbation'. It was
reported that gravity waves provide the seed mechanism for creating large-scale irregularities [Whitehead, 1971]. It was proposed that gravity waves can produce strong ionization perturbations when the phase speed of a gravity wave is equal to the drift speed of ionization, which was termed the spatial resonance. Huang and Kelly [1996a] have extensively studied the properties of large-scale equatorial F-region irregularities produced by gravity waves, using simulations. They have separated the different processes such as spatial resonance and R-T instability to show their relative importance. It was concluded that the R-T instability is the most important mechanism for the production and rise of the plasma bubble structures. Fagundes et al. [1999] also showed the importance of waves as 'seeding sources' for the occurrence of ESF. Fejer et al., [1999], using the Jicamarca incoherent scatter radar observations during 1968-1992, concluded that when the drift velocities were large enough, the necessary seeding mechanisms for the generation of ESF always appears to be present. All these studies, even though differ regarding the importance of seeding perturbations, show that the background ionospheric and thermospheric conditions are the crucial factors for triggering the ESF.

It is clear that all the aforementioned intriguing phenomena are essentially the outcome of the complex electrodynamical as well as neutral-dynamical processes characteristic of the low-latitude region. The present investigation was carried out in order to understand some of these phenomena, like EIA, ESF and their interrelationships. In the next section, the important scientific objectives of this study are mentioned briefly.

1.10 Present Study

It is evident from the discussion on the EITS processes (namely the EIA, ETWA and ESF) that these are in fact different manifestations of the ionosphere-thermosphere coupling over low and equatorial latitudes. However, even establishing a qualitative relationship among them has taken decades. For instance, the EIA was discovered as early as 1946, while presence of ETWA was indicated only in the 1990s and our comprehension about their interrelationship has evolved only recently. These processes,
especially the ESF and ETWA cannot be predicted using the present day ionospheric-thermospheric models. This indicates that the quantitative understanding of the equatorial ionospheric-thermospheric processes is still incomplete and the existing models need to be improved for better representation of the EITS. In recent times, the need to better understand and model the equatorial ionospheric processes has also been felt while addressing the satellite based navigation issues, such as Global Positioning System (GPS) based navigation. The present thesis highlights the importance of these processes i.e., EIA, ETWA and ESF in this context.

The major emphasis of the present work is on studying the spatial and temporal variability of the ionospheric density over low-and equatorial latitudes using the Computerized Ionospheric Tomography (CIT) technique. The work presented in this thesis is based on the TEC measurements obtained using the Indian Coherent Radio Beacon Experiment (CRABEX). This experiment was initiated during the year 2001-2002. The details of this experiment are described in the following chapter. The tomographic images presented here are the first of its kind from the Indian longitudes. While these images clearly bring out some unique features of the EIA and ESF, the studies using single station TEC measurements have led to some important conclusions concerning the F₃ layer/topside ledge and the ESF/Scintillations and its deterministic prediction. All these aspects are discussed in detail in the following chapters.