Chapter - I
INTRODUCTION

1.1 Classification of the earth's atmosphere

The earth’s atmosphere can be classified into different regions according to the dominant physical and chemical processes or properties prevailing at various altitudes. Based on the composition of major constituents, it is divided into two -- homosphere (or turbosphere) below ~100 km and heterosphere above. Homosphere is the well mixed part of the atmosphere, where the relative concentrations of major constituents are kept the same throughout the altitude range (irrespective of the change in pressure) by turbulence. Above turbopause, which is the upper boundary of the homosphere, the atmospheric constituents are under hydrostatic equilibrium with each species having its own scale height.

Another classification of the atmosphere is based on the temperature gradients existing at various altitudes. The earth’s atmosphere is primarily heated in three regions by three different processes. In the lower most region of the atmosphere, the triatomic molecules such as water vapour and carbon dioxide absorb mainly the longer wavelength (the infrared, IR) radiation. Earth being the primary source of this radiation, there exists a negative gradient of temperature with altitude near the surface and this region is termed as troposphere. The second temperature maximum is maintained by ozone, the vertical distribution of which extends, roughly, between 10 and 80 km with peak concentration at about 25 km. Ozone molecules absorb solar
ultra violet (UV) radiation and the maximum temperature occurs at a level (~50 km) where the concentration of ozone and intensity of UV radiation are optimum. Temperature decreases downwards and upwards from this level. Thus there exists a region of minimum temperature between this level and the earth’s surface. This level being at the top of the troposphere, is designated as tropopause. Above tropopause the temperature gradient is positive (upwards) which maintain the region stably stratified and hence this region is labelled as stratosphere. The level at which this second temperature peak occurs is at the top of the stratosphere and hence the name stratopause. Above stratopause temperature again decreases with height (this region is the mesosphere) to a minimum value (~180 °K) at mesopause.

Thermosphere is the region above mesopause where temperature gradient remains positive. In this region solar extreme ultra violet (EUV) and X - radiations dissociate molecular oxygen and ionize nascent oxygen, producing very high temperatures (1000-2000 °K) and ionization. The temperature increases to an asymptotic value determined by the variable source of solar heating. The upper boundary of the thermosphere, which is poorly defined, is around 600 - 800 km, where it merges with the exosphere, a region where collision frequency between the particles is so low that the neutral particles move in ballistic orbits subject only to gravity.

The earth’s atmosphere is also divided based on the ionization content and its influence on the propagation of radio waves. The ionosphere is defined as that part of the atmosphere, where sufficient quantities of ionization exist to influence the propagation of radio waves through it. The ionosphere is again classified depending on its ionization number density into D, E and F regions.
1.1.1 D Region

This is the lower most part covering the altitude range of 60-85 km of the ionosphere; height of peak electron density occurs at ~75 km (not well defined). Ion concentration in this region is mainly produced by the ionization of Nitric Oxide (NO) by the Lyman-α line (of the solar spectrum) and that of oxygen molecules by EUV or X rays. Collision frequency of electrons in this region with neutral particles being very high due to the latter’s abundance, this region causes strong attenuation of radio waves. Daytime electron density in this region is of the order of $10^3$ cm$^{-3}$ in normal conditions, but it increases by as much as 2 orders or even more when an intense solar flare occurs.

1.1.2 E region

This region covers the altitude range of about 85 to 140 km of the ionosphere. Maximum electron density of the E region generally occurs near ~105 km. Ionization in this region arises due to the absorption of EUV between 800 and 1026 Å and X-rays from 10 to 100 Å. The most abundant ions in this region are found to be NO$^+$ and O$_2^+$. E region covers the altitude range between the levels where $\omega_i = \nu_i$ and $\omega_e = \nu_e$ (where $\nu$ is the collision frequency of charged particle with neutrals and $\omega$ is the gyrofrequency and subscripts $e$ and $i$ denotes quantities corresponding for electrons and ions respectively) and resulting in dissimilar mobilities of ions and electrons. Because of this, both electric field and wind produce drift velocities inclined to themselves, different for either species, giving rise to a unique situation that exists in the E region.
1.1.2.1  **E region dynamo.**

The existence of a conducting layer and the magnetic field in conjunction with the winds of tidal origin results in dynamo action producing electric fields of sufficient magnitudes in the E region to cause important effects in the ionosphere. At the mid and high latitudes, the tidal winds and the downward component of magnetic field create differential movements of ions and electrons setting up electric fields of global nature. The electric currents due to this dynamo action are of sufficient strength to cause magnetic field variations on the ground.

1.1.3  **F region.**

This is the top region of the ionosphere. The level of peak ionization generally lies in the range 300 - 450 km. The most abundant ions in this region are O\(^+\) and neutrals are O, N\(_2\) and O\(_2\). Often, during daytime, F region is bifurcated into F\(_1\) and F\(_2\) basically because of the difference in altitude variation of rate of production and loss of ionization (Ratcliffe, 1956).

At the F region altitudes the collision frequency of charged particles with neutrals is much less compared to their gyrofrequencies resulting in large values of Hall mobilities compared to Pedersen mobilities for an applied electric field. Since the force acted upon by a charged particle due to an electric field depends only on the amount of charge and the strength of the field, an electric field drives both ions and electrons in a direction perpendicular to both the electric field and magnetic field with the same velocity, constituting a plasma drift. Longitudinal (parallel to magnetic field) mobilities (of both ions and electrons) being very high at the F region altitudes, even low values of collision frequency between neutrals and ions produce ion drifts along the field lines with velocities virtually equal to the component of
neutral wind velocity in the direction of magnetic field line. Since electrons are tied to ions by electrostatic force of attraction, this results in a plasma motion parallel to the magnetic field line. Diffusion of charged particles along the magnetic field line also is very significant, especially at higher altitudes of F region, owing to the longitudinal (along magnetic field lines) mobility at these altitudes and the greater time constants of ionization loss reactions. Throughout the ionosphere, the longitudinal conductivity is so large (owing to the large longitudinal mobilities of electrons and ions) that it is reasonable to assume magnetic field lines to be electrically equipotentials. Electric fields can thus be mapped between various regions of ionosphere / magnetosphere. The electric fields generated by the E region dynamo are thus mapped into the F region where they cause electrodynamic Hall drifts (plasma). Thus the E region is considered of acting as a dynamo driving the F region load (Martyn, 1953).

1.2 Equatorial Ionosphere

Due to the magnetic field line geometry, magnetic equatorial (or low) latitude E and F regions exhibit very interesting phenomena/anomalies. Equatorial Electrojet, Equatorial Ionization Anomaly and Post Sunset Enhancement in the altitude of F region are some of them.

Global zonal electric field (due to E region dynamo) in conjunction with the horizontal magnetic field creates vertical polarization fields at the equatorial E region (since Hall mobility of electrons is much greater than that of ions for an applied electric field at E region altitudes). These polarization fields can drive large amounts of current through Hall drift of electrons in the zonal direction. The vertical polarization fields created by the magnetic field line geometry thus manifests as an enhanced Pedersen conductivity in the equatorial E region. The resulting intense and localized stream of electric...
current is usually known as Equatorial Electrojet (EEJ). Under normal conditions the direction of this current is eastward during day and westward during night. As the vertical polarization field can be sustained only by more or less horizontal magnetic field lines, EEJ is confined to a region a few degrees in width (north-south) about the magnetic equator. The latitudinal extent of EEJ is found to be ±3° geomagnetic latitude.

The F region at low latitudes is very peculiar in that, sometimes, electron concentrations are greater at midnight than at noon. Peak electron density in the F region as a function of latitude shows a pronounced 'trough' centered on the magnetic dip equator with crests at 15° to 20° north and south (Appleton, 1946). The trough is also found in the number density of electrons at fixed heights below (Croom et al., 1959) and above (King et al., 1964) the F₂ peak. This phenomenon is known as equatorial anomaly (or Appleton anomaly). This anomaly exists during most of the day and is most pronounced around sunset, but disappears after midnight (Rastogi, 1959). It is known to be caused by the electrodynamic vertical drift of F region plasma followed by plasma diffusion along the magnetic field lines to form the so called equatorial fountain (Mitra, 1946; Martyn, 1947; Duncan, 1960). At most of the times except during equinoxes, the equatorial trough is found to be asymmetrical about the equator, to a greater extent than can be attributed to variations of solar zenith angle. Hanson and Moffett (1966) showed that asymmetries could be produced by transequatorial meridional winds of the order of 50 m/s.

Thus the dynamics of the F region is affected by both electric field and neutral winds. In addition, diffusion of ionization along the magnetic field lines also play a major role in the ionization distribution. The collisions between the neutral and ionized particles result in momentum transfer between them and thereby leading to coupled equations for their motion, which makes the situation even more complicated (Rishbeth, 1972).
Plasma transport plays an important role in the theory of the ionospheric F region. The lower portion of the F region is known to be dominated by photochemical processes and the upper portion is dominated by horizontal and vertical motions of the ionospheric plasma. These motions are a combination of diffusion along the magnetic field lines, $\mathbf{E} \times \mathbf{B}$ ($\mathbf{E}$ is the electric field and $\mathbf{B}$ is the magnetic induction) drifts perpendicular to field lines and neutral winds pushing the ionization along the field lines. Near the magnetic dip equator, however, vertical plasma motion is purely due to electromagnetic drift as the field (magnetic) aligned plasma flow has no component in the vertical direction (since the magnetic field lines are horizontal) and it is thus possible to derive electric fields from this vertical plasma drift.

A full understanding of the earth's thermosphere requires a knowledge of its mean compositional, dynamical and thermodynamic state as a function of position, time, solar cycle and geomagnetic activity level. It further requires an understanding of the major physical and chemical processes responsible for the thermosphere structure and of the characteristics and causal mechanisms for the natural variability that occurs in the upper atmosphere. Also it requires a quantitative understanding of the role played by the thermosphere as a modifier and transformer of energy and of the nature of the physical mechanisms that couple the thermosphere to the mesosphere below and to the magnetosphere above.

1.3 **Electrodynamics of the equatorial F region**

Equatorial and low latitudes F region electrodynamic drifts (both vertical and zonal) have been studied by several workers using different techniques during the last three decades. Vertical plasma drifts of F region have been studied using incoherent scatter radars (ISR) (e.g., Woodman and...
Hagfors, 1969; Woodman, 1970; Balsley, 1973; Behnke and Harper, 1973 Burnside et al., 1983b), HF Doppler radars (e.g., Balan et al., 1982, 1992), Phase path techniques (e.g., Krishna Murthy et al., 1976: Subbarao and Krishna Murthy, 1983, 1994) and χF values from ionosonde (e.g., Abdu et al., 1981a, b, 1983; Nelson et al., 1986). In addition to these ground based measurements, the in-situ vector measurements of electric field have been carried out using the Ion Drift Meter (IDM) and Vector Electric Field Instrument (VEFI) on board the Dynamic Explorer-2 (DE-2) and Atmospheric Explorer-E (AE-E) space crafts (Coley and Heelis, 1989; Coley et al., 1990). The F region east-west drifts have been studied with incoherent scatter radar (ISR) observations (Woodman, 1972; Fejer et al., 1981), spaced receiver observations (e.g., Chandra et al., 1970), radar interferometer measurements (Kudeki et al., 1981), all-sky photometric imaging of field aligned (magnetic) air glow depletions (Weber et al., 1978; Mendillo and Baumgardner, 1982) and VEFI/IDM on board DE-2/AE-E (Maynard et al., 1988; Aggson et al., 1987).

These observations have determined the general characteristics of F region vertical drift and its dependence on season, solar activity, and magnetic activity. The vertically upward drift of F region ionization observed during daytime in the equatorial (magnetic) region confirmed the fountain theory of equatorial anomaly (Hanson and Moffett, 1966; Sterling et al.. 1969). Altitude profiles of vertical drift obtained using ISR showed that the vertical drift does not vary much with altitude (Woodman, 1970; Pingree and Fejer, 1987). It may be noted here that the F region vertical plasma drift at the dip equator is a measure of the zonal electric fields which are linked by magnetic field lines to the E region at a higher latitude (still in the low latitude / equatorial region). The general behaviour of the vertical drift shows an upward drift by day and downward by night, the time of reversal of the direction of drift being after sunset. This is in general agreement with the direction of global E region dynamo electric fields - eastward during day time and westward during nighttime. An unfailing observation of the vertical plasma drift of the
equatorial F region has been the enhancement of the velocity (upward) after sunset but before its reversal to the nighttime downward drift. This enhancement in the F region upward velocity or the eastward electric field is commonly referred to as the post-sunset enhancement or pre-reversal enhancement. The amplitude of this pre-reversal velocity enhancement shows large day-to-day variability even during geomagnetically quiet periods. The pre-reversal enhancements of the Jicamarca upward drifts increase significantly with solar activity during equinox and summer, but saturate for large values of the 10.7 cm solar flux ($F_{10.7} > 180 \times 10^{-22} \text{Wm}^{-2}\text{Hz}^{-1}$) during the June solstice (Fejer et al., 1989, 1991). The general dependence of the F region upward drifts on solar flux was also studied from ionosonde (Abdu et al., 1981b; Goel et al., 1990) and HF radar measurements (e.g., Namboothiri et al., 1989). Deminov et al. (1988) used $f_{\circ}F_2$ observations from Interkosmos-19 satellite to calculate the longitudinal variations of the equatorial F-region vertical velocities.

A number of theoretical (numerical) models of global or equatorial dynamo electric fields and currents have been developed (e.g., Richmond et al., 1976; Stening, 1981). Many studies have been made on the importance of tidal winds of E region origin and below in producing the observed daily Sq variation and the F region vertical drift (Tarpley, 1970; Richmond et al., 1976; Forbes and Lindzen, 1977). The diurnal trapped tidal mode (1,-2) is found to be the principal contributor to the ionospheric Sq currents and electric fields (Stening, 1969; Tarpley, 1970) and has only slight phase change between E and F regions (Volland and Mayr, 1972). This mode generated by the in-situ absorption of UV and EUV radiation is most important between 120 and 200 km during the day. Richmond et al. (1976) showed that the addition of the semidiurnal (2,4) mode to the diurnal mode modified by ion drag brings the theoretical results in closer agreement with low and mid latitude experimental drift data. The semidiurnal mode is highly variable in time and this could
explain partly the variability of the observed drift data. Richmond (1979) presented a review of dynamo theories. Additional tidal modes are likely to be important in explaining the world wide system of currents and electric fields (Forbes and Lindzen, 1976a, b, 1977; Stening, 1977). Stening’s calculation also shows that some aspects of the F region drift are consistent with the calculated E region polarization fields during daytime but not at night. In particular the agreement between his calculation and the observed eastward field at the equator is poor except around midday. Forbes and Lindzen (1977) suggested that the disagreement is caused most likely by the neglect of the F region dynamo system (Rishbeth, 1971a, b, 1981).

While the Sq current system is well predicted by the E region dynamo theory, the daily variations of the vertical plasma drift in F region are not well reproduced except for the very general behaviour of day time upward drift and nighttime downward drift. Discrepancies are found between observation and model predictions (E region dynamo) in the reversal time of drift velocities after sunset and nighttime in general. The above mentioned models which take into account only the tidal modes of E region to calculate the E region dynamo failed to predict the characteristics of post sunset enhancement of the vertical plasma drift. It is now fairly accepted that the discrepancy is caused by the non-inclusion of F region dynamo (polarization) fields.

1.3.1 F region dynamo

Rishbeth (1971a) first put forward the idea of F region having its own dynamo proper. Thermospheric winds (meridional component) at F region altitudes induce a plasma drift along the magnetic field line, as already explained. In addition there can be a much slower drift perpendicular to the wind and magnetic field line. This arises from the fact that the Hall mobility
of ions for a wind induced mechanical force is large compared to that of electrons at the F region altitudes (since the frequency of ion-neutral collisions is greater than that of electron-neutral collisions). Nevertheless, wind can also cause the electrons to drift but in opposite direction to that of ions and at a very low speed since electrons are more bound to the magnetic field line due to their greater gyrofrequency. Thus the Hall drift of ions induced by wind constitutes an electric current. As the current is generated from the action of wind blowing across magnetic field lines, it was designated by Rishbeth (1971a) as F region dynamo in par with the E region dynamo.

Rishbeth (1971b) discussed the role of E region conductivity in developing the F region polarization fields. At the equatorial (magnetic) latitudes an eastward wind can generate an upward ion drift by the dynamo action as already explained. This vertical ion drift can only be maintained by a Pedersen current through the E region (in magnetic meridional direction). Since at the top of the E region Pedersen mobility of ions is larger compared to that of electrons, the E region current is mainly constituted by ions. The F region vertical current is linked to the E region by an electron flow via the magnetic field lines constituting a field aligned current. When the E region is sunlit it almost entirely short circuits the F region polarization field. After sunset, however, E region conductivity may be too small to support the field aligned currents and as a result vertical polarization field develops at the F region. At equilibrium this polarization field balances the dynamo field preventing further vertical ion flow. This polarization field maintained by an eastward wind accumulates positive ions at the top of the F region and negative charges at the bottom. This charge density is proportional to the wind velocity. This vertical polarization field explains the zonal plasma drift enhancements during nighttime (Rishbeth, 1971b; Heelis et al., 1974). It may be noted here that these field aligned currents can also be produced by the asymmetry of the conductivities in the two hemispheres which may flow from
one location of E region (off-equational) to the linked E region of the other hemisphere through the magnetic field line (Walton and Bowhill, 1979).

Besides F region polarization fields just described, there are polarization fields of E region origin. These fields, suggested by Martyn (1953) are produced by the dynamo action in the E region by tidal winds which exist in various diurnal, semidiurnal etc. modes as mentioned earlier. These fields are of similar magnitudes by day and by night, unlike F region polarization fields which build up quickly at sunset and decrease quickly at sunrise. Thus F region may play a greater role than the E region in the F region nighttime phenomena.

Heelis et al. (1974) studied in detail the effect of F region dynamo in modifying the F region vertical drifts which would otherwise be driven by E region electric fields. They considered the system of E and F regions as separate slabs of appropriate thicknesses in which electrical coupling is provided through the highly conducting magnetic field lines connecting them. The procedure by which the numerical calculations were carried out is as follows. First, the E region currents and electric fields are calculated as driven by the E region tidal winds, assuming field aligned current to be zero. Then from the F region winds and the electrostatic fields mapped up from the E region, the modified wind and motion of F region plasma are calculated and also the small ion current flowing perpendicular to wind and the magnetic field line (F region dynamo) is calculated. The field aligned current required to satisfy the divergence free nature of this ion flow is obtained. E region current and fields are then recalculated taking into account this additional current source (or sink). This modified electrostatic field is again fed back into the F region equations and the whole process is repeated until it converges to a self-consistent steady state.
Any convergence or divergence of current flow in the F region leads to a set up of polarization fields but this can be discharged by currents flowing along the field lines and through the E region. As a result polarization fields depend on the conductivity of the E region and hence the polarization fields tend to be large at night when E region is poorly conducting and small by day when E region is highly conducting.

Heelis et al. (1974), using the above mentioned model, found that the inclusion of the field aligned currents to take account of the F region polarization fields has quite a dramatic effect on the vertical plasma drift velocity in the F region. During the day, the field aligned current changes the phase and maximum daytime value of the drift velocity in the F region. At the time of sunset the conductivity of the E region suddenly decreases and the large field aligned currents at this time cause a large longitudinal gradient in the electrostatic potential in the E region and consequent sharp increase in the F region plasma drift velocity. During nighttime, when the E region conductivity is small the E region electric field and the F region plasma drift are very sensitive to changes in field aligned current. Their model demonstrated that the inclusion of F region dynamo reproduces many features of the observed F region vertical drifts.

Burnside et al. (1983b) examined the importance of F region polarization electric fields over Arecibo (geog. lat. 18.6° N, long. 66.8° W; dip 50° N) using nighttime measurements of E and F region electrical conductivities and of ion drift and wind velocities. They showed that during solar maximum the height integrated F region Pedersen conductivities exceed the E region conductivity on the average by a factor of about 10 throughout the night. As a result, the Arecibo nighttime F region electrodynamic drifts are best explained by including the effects of F region polarization electric fields. Walton and Bowhill (1979) explained the post sunset enhancement without
considering F region dynamo. But they included field aligned currents in their calculation by not assuming the symmetry between the hemispheres and by taking into account the non-coincidence of magnetic and geographic equators and they argued that for Jicamarca, where this difference is large, F region dynamo need not be included to explain the post sunset enhancement.

Following the model developed by Heelis et al. (1974), Farley et al. (1986) performed model calculations of equatorial electric fields and discussed a physical mechanism by which F region vertical polarization fields result in the enhancement of zonal electric fields (post sunset enhancement), in terms of longitudinal gradient of E region conductivity. A zonal neutral wind \((U)\) can generate equatorial vertical polarization (electric) field of \(U \times B\) (direction being opposite to that of normal ion flow, and is vertically downward, if \(U\) is eastward). This electric field is partially short circuited by the E region depending on the relative values of the field line integrated conductivities of E and F regions. They showed that the evening enhancement of the eastward electric field is a direct consequence of the considerably more rapid decrease of E region conductivity compared to that of F region. Vertical polarization (downward) electric field set up at the F region altitudes is mapped to the E region as a meridional (equatorward) electric field through the highly conducting magnetic field lines. During daytime this field is shorted out by an ion flow in its direction at the top of the E region. During evening hours, after sunset, the Pedersen conductivity of the E region abruptly falls to its nighttime value and fails to short circuit this field. This equatorward field (for an eastward wind at F region altitudes) drives an eastward electron flow through Hall drift. At this time a few degrees eastward (in longitude), the E region is already in darkness and conductivity will be too weak to support this electron drift, resulting in an accumulation of electrons at the dusk terminator. This gives rise to an eastward electric field at the time of sunset and a westward field after some time (or few degrees eastward).
This eastward (and then westward) electric field is mapped back into the F region where it causes upward (and then downward) drifts as observed. The day-to-day, seasonal and solar cycle variations of this pre-reversal enhancement could be attributed to the variations in the F region wind changes, the ratio of the E and F region field line integrated conductivities and to the spherical asymmetries arising out of the non-coincidence of geographic and magnetic equators.

Maeda et al. (1982) observed a distinct variation in the geomagnetic D-component near the dip equator using the NASA Magnetic Field Satellite (MAGSAT). The amplitude of D-component depends on geographic longitude such that it is large in American zone and small in Indian zone. It is found to be larger during high solar activity periods. Takeda and Maeda (1983) explained this anomalous variation of magnetic declination and its dependence on solar activity on the basis of meridional currents (electric) in the F region caused by neutral pressure gradient, supporting the theory of F region dynamo.

Tsunoda (1985) and Batista et al. (1986) have linked the seasonal and longitudinal variations of the equatorial scintillation onset and post sunset enhancement with the longitudinal variation of the integrated E region Pedersen conductivity due to the offset between magnetic flux tubes and the evening solar terminator.

Goel et al. (1990) attributed the solar cycle dependence of the post sunset enhancement to the variation in E region conductivity, which is greater during solar maximum period (by a factor of 2) than during solar minimum period. Takeda and Araki (1985) showed that the nighttime height integrated Pedersen conductivities also exhibit large solar cycle variation due to change in electron density and ion-neutral collision frequency in the lower ionosphere.
Haerendel and Eccles (1992) studied the role of equatorial electrojet (EEJ) in the evening ionosphere. They suggested that the equatorial electric field in the evening sector results from a large current system set up by the effects of the F region neutral wind dynamo and equatorial electrojet. This current is upward at the equator since the upward current driven by the F region dynamo is not balanced by the Pedersen current (Haerendel et al., 1992). They suggested that the EEJ plays an important role in the evening enhancement of upward and eastward plasma drifts.

Crain et al. (1993b) constructed a model in which the plasma distribution and the dynamo generated potential distribution are self-consistently calculated. In this model, E and F regions are considered as a single system without making distinction between different regions and currents are allowed to be generated and closed in a manner consistent with the conductivities and wind distributions, representing a true closed circuit model (e.g., Richmond and Roble, 1987; Singh and Cole, 1987a, b, c). They suggested that the F region plays a greater role in the low latitude dynamo than previous models have predicted. Though the peak value of Pedersen conductivity in the E region is greater than the peak value of the F region Pedersen conductivity, the field line integrated Pedersen conductivity of the F region can be significantly greater than the integrated Pedersen conductivity of the E region. This effect is most pronounced for low latitude flux tubes with apex altitudes slightly above the F region peak. This arises due to the geometry of the equatorial geomagnetic field lines in which the length of magnetic field line through F region is much greater than the length through the E region. This effect decreases with increasing latitude as the field line becomes more vertical. Thus F region may provide a significant contribution to the dynamo electric fields at all local times in a limited altitude and latitude region (Crain et al., 1993a).
1.4 Neutral winds in the F region

Neutral wind is a very important thermospheric parameter which significantly influences the distribution of F region ionization and its peak density through transport of ionization and various other interrelated processes. Thermospheric winds transport energy and momentum between various regions especially during geomagnetic storms. The interaction of F region plasma and the neutral winds is due to the collisions between neutral particles and ions resulting in a frictional force. This causes plasma to be driven in the direction of geomagnetic field line at a speed equal to the component of wind along the line. If the dip angle of the field line is \( I \), a wind of velocity \( U \) in the direction of magnetic meridian can impart a vertical plasma drift of \( U \cos(I) \sin(I) \) (Rishbeth and Gariott, 1969). A transfer of momentum is accompanied by a transfer of energy and thus the neutral winds not only drift plasma but also heats the ionosphere (Stubbe and Chandra, 1971). In addition, neutral dynamics affects the heat balance of thermosphere through transport processes involving radiatively active trace constituents such as Nitric Oxide. Thermospheric winds also play an important role in the F region behaviour during magnetic disturbances (Rishbeth, 1975) especially in bringing about a change in the chemical composition (Mayr and Volland, 1973) and through the disturbance dynamo electric fields (Blanc and Richmond, 1980). Compositional changes in the neutral atmosphere affecting the atom/molecule concentration ratio would affect the loss coefficient.

Measurement of thermospheric winds at middle and low latitudes is important for an understanding of the mean circulation as well as the propagation characteristics of waves and perturbations originating at high latitudes. As the thermospheric neutral winds affect the formation and growth of irregularities, neutral wind plays an important role in the equatorial phenomena such as spread-F, airglow depletions, ionospheric bubbles and
ionospheric scintillations, which are basically generated by Rayleigh-Taylor
instability (See Meriwether et al., 1986; Mendillo et al., 1992). Maruyama and
Matuura (1984) attributed the onset of ionospheric scintillations during
equinoctial periods to the small magnitude of the transequatorial wind at this
time, and the lack of scintillation activity during winter solstice to the
asymmetric distribution of ionization along the field line caused by
transequatorial winds (which is usually directed from summer to winter
hemisphere). Aburr-Robb and Windle (1969) studied the effect of meridional
neutral wind on the redistribution of ionization in the longitude direction
affecting the equatorial ionization anomaly or Appleton anomaly (See Rishbeth,
1972 for a review).

The earth’s upper atmosphere has been extensively studied in the
late 1950s by means of its effects on the orbits of artificial satellites. Accurate
orbital observations have provided extensive information on the air density in
the thermosphere (King-Hele, 1992). The existence of thermospheric neutral
winds of sufficient magnitudes to influence the dynamical behaviour of the F
region were known from the early 1960s. King and Kohl (1965) for the first
time made calculations to establish the importance of neutral winds on
ionospheric drifts at F region altitudes. It was known from the studies of
orbital changes of artificial satellites, as mentioned above, that the
atmospheric pressure at F region altitudes varies as a function of latitude and
local time (King-Hele, 1959). Thermal expansion caused by the solar heating
of the atmosphere forms a high pressure bulge on the dayside giving rise to
horizontal gradients of air pressure which can drive winds. This knowledge
of existence of day-to-night atmospheric pressure difference at the F region
altitudes led Geisler (1966, 1967) and Kohl and King (1967) to attempt detailed
calculation of the neutral wind system that should result. These early models
predicted mainly diurnal winds blowing from a pressure bulge on the dayside
to a pressure minimum on the night side.
Movement of the ionized particles being constrained along the magnetic field lines, the ionosphere offers a frictional force (ion drag) to the neutral winds driven by the pressure gradients. As the time scales of damping caused by ion drag are small compared to the time scales of Coriolis or inertial terms, the winds at F region heights tend to blow virtually in the direction of pressure gradient force, i.e., in the great circle paths, directly from the hottest part of the upper atmosphere (near subsolar point) to the coolest region in the dark hemisphere (near the antisolar point) (Dickinson et al., 1975). Though not strong, Coriolis force also affects the thermospheric winds. As a result the direction of wind will be inclined to the pressure gradient force depending on the relative magnitudes of ion drag and Coriolis terms. Apart from these forces, the winds are also affected by viscosity of the air arising out of the greater mean free path of atmospheric constituents. Since the kinematic viscosity is inversely proportional to the density of the gas, its effect is more pronounced at greater altitudes. The extent to which wind speed and direction vary with altitude at any given place and time is strongly influenced by viscosity. Viscosity scale height parameter (within this altitude range viscosity smooths out vertical gradients in horizontal wind) as defined by Rishbeth and Garriot (1969) is found to be of the order of 100 km during nighttime (Rishbeth 1972).

Various experimental techniques are available to measure/deduce thermospheric neutral winds (or its components) such as vapour trail method (e.g., Haerendel et al., 1967), Fabry-Perot interferometers (e.g., Armstrong, 1969; Sipler and Biondi, 1978; Biondi et al., 1988; Meriwether et al., 1986), incoherent scatter radars (e.g., Vasseur, 1969; Salah and Holt, 1974; Harper, 1977; Burnside et al., 1991). In-situ measurements of thermospheric winds were realized using the Atmospheric Explorer-E and Dynamic Explorer-2 Satellites (Spencer et al., 1981, 1982; Killeen and Roble, 1988a, b). Miller et al. (1986) and Buonsanto (1986) deduced meridional winds from $h_m F_2$.

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data obtained using ionosondes. Burrage (1990) obtained thermospheric wind from the brightness measurements of 6300 Å night glow from AE-E Satellite. These methods are described in Chapter-IV.

Serious discrepancies with the earlier model predictions of diurnal wind pattern were evident at low latitudes. The earlier simple models predicted maximum equatorward wind in the post midnight period. However, Nelson and Cogger (1971) showed that the heights of the F layer typically decreased by 50 km after midnight at Arecibo. Wright (1971) used spaced receiver technique to conclude that the decrease in the F layer height observed at Conception, Chile (geog. lat. 35°S) was due to a variation in the neutral wind. He also noted that the decrease in the height of the F layer tends to maximise near a magnetic dip angle of 45° at both the hemispheres, stressing the idea that the depression is caused by a change in neutral wind rather than zonal electric field. Behnke and Harper (1973) and Harper (1973) interpreted this descent of the F layer in terms of the decrease and frequent reversal of the meridional wind. Now, it is fairly established that this reversal is caused by an enhancement in thermospheric temperature during midnight hours. This phenomena exhibited by low latitude thermosphere is termed as Midnight Temperature Maximum or MTM (e.g., Herrero 1993, Sastri et al., 1994).

Harper (1973) suggested that a semidiurnal component of wind can cause this anomalous temperature maximum during midnight hours. Harper (1979) by analyzing meridional wind data obtained using Arecibo incoherent scatter radar showed that a semidiurnal tide is significant in meridional wind variation at F region altitudes. This is in agreement with the predictions of Hong and Lindzen (1976) and Garrette and Forbes (1978). Spencer et al. (1979) from in-situ measurements of neutral temperature using NATE (Neutral Atmosphere and Temperature Experiment) instrument of AE-E.
satellite, established the existence of a temperature maximum around midnight and suggested that momentum coupling associated with ion drag and tidal waves from the lower atmosphere should be a likely source mechanism for generating MTM. Mayr et al. (1979) described these processes in the context of tidal theory. They attributed MTM to the generation of semidiurnal and terdiurnal tides from non-linear interaction between the diurnal variation of wind and diurnal variation of ion density.

Though considerable progress has been made in atmospheric tidal theory (e.g., Forbes 1982a, b; Fesen et al., 1991a, b) over the past two decades, actual tidal decomposition studies performed on experimental thermospheric data are very few due to the sparsity of data, especially at low/equatorial latitudes, before satellite measurements were available. Buonsanto (1991) performed a tidal decomposition on the meridional wind data obtained over a full solar cycle at Wallops Islands (geog. lat. 37.8° N, long. 284.5° E) using the servo model method of Buonsanto (1986). He found significant amplitudes for diurnal, semidiurnal, terdiurnal and quarterdiurnal components. His results are in agreement with that of Amayenc (1974) for St.Santin (geog. lat. 44.6°N, long., dip 2.2°E). Mayr et al. (1979) and Herrero et al. (1983) performed tidal decomposition on the thermospheric meridional wind data obtained using NATE instrument on board Atmospheric Explorer-E satellite and explained midnight temperature maximum on the basis of tidal theory.

1.5 Thermosphere models.

Many models of the thermosphere parameters have been developed in the past three decades. These range from simple empirical specifications of one or two thermosphere parameters to highly complex 3-dimensional numerical models.
MSIS models of Hedin (1983, 1987) are empirical models of thermospheric temperature, density and composition based on the in-situ data from satellites, as well as incoherent scatter radars. MSIS-86 of Hedin (1987) (stands for Mass Spectrometer and Incoherent Scatter) model is a refined version of the earlier model MSIS-83 (Hedin, 1983) with improved representation of polar region morphology. This model is particularly valid for altitudes above 150 km as the global data coverage for density and temperature and particularly composition are very sparse between 90 and 150 km. So, the model predictions for this altitude range are not expected to be as representative as at higher altitudes of the atmosphere. So is the case with exosphere above 600 km where extrapolations are expected to become increasingly inaccurate at higher altitudes.

In recent years, Thermospheric General Circulation Models (TGCMS) have been developed to study the global circulation pattern, temperature and compositional structure of the atmosphere. In particular, two numerical general circulation models - the National Center for Atmospheric Research (NCAR) TGCM and the University College London (UCL) TGCM have been successful in calculating wind and temperature fields similar to those observed from DE-2. NCAR-TGCM is based on the models developed by Dickinson et al. (1981) and Roble et al. (1982, 1983, 1984) and UCL-TGCM is based on the global model of Fuller-Rowell and Rees (1980). The main inputs for TGCM are solar EUV/UV fluxes, high latitude forcing and upward propagating tides. Roble et al. (1987) developed a global average (1-D) model of the coupled Thermosphere and Ionosphere General Circulation Model (TIGCM) (Roble et al., 1988).

Killeen (1987) developed a model based on NCAR-TGCM outputs -- the VSH (Vector Spherical Harmonic) model. VSH model is basically a wind and temperature synthesizer relying on the outputs from sample runs of the
NCAR-TGCM. The model is based on a spherical harmonic expansion of outputs from NCAR-TGCM. Thus it generates model values similar to that of NCAR-TGCM without the use of a super computer. VSH model contains a description of wind field using Vector Spherical Harmonic expansion in universal time and a polynomial expansion in altitude.

Hedin et al. (1988) developed an empirical model analogous to MSIS models, based on the wind data obtained from Atmospheric Explorer-E and Dynamic Explorer-2 satellites. This horizontal wind model (HWM-87) is valid for altitudes above about 220 km where experimental data coverage is best and altitude variation of horizontal velocity are reduced by viscosity. This model does not predict solar activity effect but does predict magnetic activity effects. HWM-90 (Hedin et al., 1991) is a revised version of HWM-87. In HWM-90 AE-E and DE-2 data are supplemented by wind data for lower and upper thermosphere from ground based incoherent scatter radar and Fabry-Perot optical Interferometers. The ground based data allow modelling of seasonal/diurnal variations. While solar activity variations are included in HWM-90 they are found to be small and not always clearly represented by the data used. The model describes the transition from predominantly diurnal variation in the upper atmosphere to semidiurnal variations in the lower thermosphere. Transequatorial winds flow from summer to winter hemisphere above 140 km while the flow is opposite for altitudes below. Also the model shows significant altitude variations up to 300 km at some local times.

In the above, a brief account of the various F region dynamical phenomena including the prereversal enhancement of the vertical drift and thermospheric winds is presented. It is evident from this account, that there are some important gaps in the understating of these phenomena. The author has carried out detailed studies on the equatorial nighttime F region vertical drifts using ionogram data. A method of deducing the equatorial nighttime
thermospheric meridional winds from ionograms has been developed. Using this method, the characteristics of the equatorial thermospheric meridional winds (nighttime) have been studied in detail. The results of these are presented and discussed in the following chapters.