Chapter 1. Introduction

This thesis is a compilation of the airglow observations conducted from Kavalur (12.5° N, 78.8° E, 4.6° N Magnetic) and Mt. Abu (24.5° N, 72.7° E; 18.5° N Magnetic) in the high solar activity period of 1999-2002. The observations are carried out using the all sky imaging system developed at Physical Research Laboratory, Ahmedabad, India. The objective of the study is to understand the characteristics of plasma depletions over the Indian region during solar maximum. Kavalur, being an equatorial station, could give features of plasma depletions in the early stages of their evolution, once the scale sizes are sufficient to be detected by the imager. From the low latitude station Mt. Abu, it is expected to study features of well developed plasma depletions. In this chapter, an attempt is made to introduce the ionospheric physics, starting with a brief description of the structure of the atmosphere and the details of the structure and formation of ionosphere. The characteristics of equatorial ionosphere and various ionospheric processes such as Equatorial Spread-F are then explained, which leads to the main topic, “plasma depletions”. The background material given below is mainly based on the works of Rishbeth and Garriott (1969), and Hargreaves (1992).

1.1 Structure of Neutral Atmosphere

The earth's atmosphere is divided mainly into four regions viz. troposphere, stratosphere, mesosphere and thermosphere, based on the vertical distribution of temperature. Troposphere, meaning turning or changing sphere, ranges from ground to about 17 km near tropics. It accounts for more than 80% of the mass and virtually all of the water vapor, clouds and precipitation in the atmosphere. All the weather systems are within the troposphere. It is characterised by very strong vertical mixing. The surface of the earth acts as a black body that absorbs the solar radiation, which is rich in short waves, and re-radiates longer wavelengths. The atmospheric molecules such as water vapour, carbon dioxide, ozone etc. absorb and emit these infrared radiations, thus providing an efficient heat transfer in the troposphere. As we go up in the troposphere, the temperature initially decreases due to adiabatic cooling, until above a certain
altitude where it remains uniform (about 17-20 km in the tropics), which is known as tropopause.

The temperature starts to increase with altitude above tropopause, and this region is known as stratosphere, which extends from 20 km to about 50 km. It is highly stratified or layered region. Unlike in the troposphere, there is very small vertical mixing in stratosphere and is characterized by the presence of ozone layer. The ozone absorbs most of the harmful ultraviolet rays from the Sun, and thus acts as heat source. This explains the observed temperature variation in the stratosphere. The region above stratosphere is termed as stratopause, where the temperature ceases to increase with altitude. The pressure at stratopause level is about 1 mb, whereas at ground it is 1000 mb. About 99.9% of the atmospheric mass resides within troposphere and stratosphere.

Mesosphere, the middle sphere, ranges from 55 km to about 85 km. In mesosphere also temperature decreases with altitude. Vertical air motions are present in mesosphere. During summer season, thin layers of clouds are produced in this region. During twilight, when lower atmosphere is in the earth's shadow, these clouds will be visible from ground as noctilucent clouds. The upper region of mesosphere, known as mesopause, falls among the coldest regions in the atmosphere. The mesospheric temperature minimum is due to the radiative cooling of that region by airglow emissions both in IR and visible range.

The region above 90 km is called thermosphere and it extends to several hundred kilometers. In thermosphere, temperature increases first steeply and then slowly with height. The temperature in these regions varies from 500 K to 2000 K. The heating is caused by the absorption of solar EUV and X-rays by the atmospheric constituents. In this process these constituents get ionised and result in the formation of ionosphere. At higher heights, the density is extremely low and heat conductivity is very high, resulting in almost uniform temperature distribution. Figure 1.1 illustrates the vertical temperature distribution in detail.

The region of atmosphere below 100 km is generally called homosphere, which is marked by turbulent mixing of the gas molecules. As a result, various atmospheric species are well mixed and are distributed according to their mean mass. In the region
above 100 km there is no turbulent mixing. Above this height diffusive separation controls the distribution of gases. This region is called heterosphere. The height region where turbulent mixing becomes less important than diffusive separation is called turbopause. Heavier elements are prominent in the lower parts and as we go up lighter elements become the important species. In the region above 600 km, where the mean free path exceeds scale height, is called exosphere. In exosphere, collisions are negligible and molecules move under the influence of gravity. If the velocity of molecules is greater than the escape velocity, they escape from the gravitational field of the earth.

Figure 1.1 Vertical structure of the ionosphere showing the different layers marked as D, E, F (F₁ and F₂) etc. This nomenclature is due to the fact that when an electrically conducting layer was first proposed to exist around 100 km, it was termed as E region. Following the convention, later when other layers were discovered to exist below and above the E region, they were respectively called D and F regions. The F₁ and F₂ regions are the sub divisions of the F region, which is found to split into two layers during day time, owing to the differences in the photochemistry at different altitudes. Also shown is the vertical distribution of temperature, based on which the atmosphere is divided into different regions as indicated. [Source: http://ion.le.ac.uk]
1.1 Structure of the ionosphere

The solar X-ray and UV radiation entering the upper atmosphere ionise the elements present at those altitudes and results in the formation of ionised layers in the upper mesospheric and lower thermospheric regions. This region is known as ionosphere. Ionosphere is defined as that region of the earth’s atmosphere, where free electrons exist in sufficient numbers to affect the radio wave propagation. Ionosphere starts from about 50 km and extends up to more than 1000 km. The presence of a conducting layer in the upper atmosphere was suggested long back in eighteenth century from the study of the geomagnetic field variations. The ionospheric studies gathered momentum in 1901, after the success of the Trans Atlantic radio propagation by Marconi. This revealed the importance of ionospheric studies in the field of radio wave communication. Various techniques were employed to study the ionosphere. They involve ground-based instruments such as conventional radar, ionosonde etc. and in-situ techniques such as Langmuir probes, double probes, etc. carried using rockets or satellites.

Ionosphere is described as a laboratory, where different verities of chemical and physical processes take place. The major source of ionisation is the solar X-ray and EUV radiation. At high latitudes the solar wind particle precipitation plays an important role in ionisation. Depending on the vertical distribution of the electron density, ionosphere is divided into different regions. They are known as D region, E region and F region etc. D region starts from about 50 km and extends to about 90 km. The important chemical species in this region are O₂, N₂ and NO. Radiation having absorption cross section less than 10⁻¹⁹ cm² causes ionisation in this region [Banks and Kockarts, 1973]. The hard X-rays (0.1- 1 nm) ionise all elements. The Ly α (121.6 nm) line ionise NO and solar UV rays (102.7-111.8 nm) ionise O₂ (¹D). Cosmic rays are also responsible for ionisation in this region. The typical electron density in this layer is about 10³ electrons per cc. The dynamics of this region is mainly determined by the dynamics of neutral elements.
The region extending from 90 km to about 140 km is called E region. The important ionising radiations in this layer are the soft X-rays (1-20 nm), which ionise all and the EUV radiation 91.1-102.7 nm that ionise O$_2$. Absorption cross section of radiation in this region is less than $10^{-18}$ cm$^2$ [Banks and Kockarts, 1973]. The prominent ions are NO$^+$ and O$_2^+$. In this region, electron density is of the order of $10^5$ cm$^{-3}$. The region above 140 km is called F region. The most important species in this layer is atomic oxygen. The ionising radiation is the EUV rays in the range 20-91.1 nm that ionise N$_2$, O$_2$ and O, and the absorption cross-section is less than $10^{-17}$ cm$^2$. F region is usually split into two layers F1 and F2 during daytime in certain seasons. The region below 50 km, where ionisation negligible compared to neutrals, is called C region. The main source of ionisation in this region is the cosmic rays that ionise all elements. In this region electron density is very small. The electron distribution with altitude is given in Figure 1.1.

1.2 Formation of the ionosphere

While the high energy radiation entering the upper atmosphere ionise the neutral atoms and molecules, the ions and electron thus produced undergo recombination reactions. Under quiescent conditions, the balance between production and loss of ions maintains the stable ionosphere. The layered structure of the ionosphere is as a result of the relative variation of the intensity of the ionising radiation and neutral distribution with altitude. Maximum ionisation occurs at the level where, the increase in the attenuation of the radiation and the gas density balances, as we go down from the top of the atmosphere. O$_2$, N$_2$ and O are the important chemical elements in the ionospheric region. These elements are ionised by the solar EUV and X-rays through photo dissociation, photo ionisation etc. and form ionised regions in the upper atmosphere. Photons with energies greater than 12 eV can ionise one or more atmospheric constituents. The processes involved in the production and loss of ions are described in detail by Chapman in 1931. The simple production function given by Chapman for a horizontally stratified atmosphere having a single component with constant scale height is,
\[
\frac{q(z)}{q_{m0}} = \exp\{1 - z - e^{-z} \sec \chi\}
\]

where, the optical depth \(z\) is given by,

\[
z = \frac{h - h_{m0}}{H}
\]

Here \(q\) is the rate of production at a given height \(h\) when the solar zenith angle is \(\chi\), \(q_{m0}\) is the maximum rate of production when the sun is overhead and \(h_{m0}\) is the corresponding height of maximum production. \(H\) is the scale height. At heights above the height of maximum production \(h_{m0}\), \(z\) is positive and for large values of \(z\),

\[
q \approx q_{m0} e^{-z}
\]

At heights well below \(h_{m0}\), \(z\) is negative and

\[
q \approx q_{m0} \exp\{-e^{-z} \sec \chi\}
\]

In the above equation, the height of unit optical depth and the height of maximum ionisation coincide, since scale height is assumed to be independent of height.

The process of ionisation is complete only if the loss mechanisms are also considered. In D region the photo-ions produced are lost through dissociative recombination with electrons or neutralisation with negative ions. The rate coefficient for the above two processes is approximately \(10^{-6}\) cm\(^3\) sec\(^{-1}\). A three-body reaction involving attachment of electrons with neutrals is suggested for the production of negative ions. Photo-detachment, associative detachment and collisional detachment are the important loss mechanisms of negative ions. The coefficient of photo detachment for O\(_2^-\) is \(\sim 0.4\) sec\(^{-1}\) and coefficient of associative attachment is of the order of \(10^{-10}\) cm\(^3\) sec\(^{-1}\).

In E and lower F regions, the important loss process is dissociative recombination. This is a two-stage process involving a charge transfer reaction followed by
dissociative recombination. The direct or radiative recombination of atomic ions and electrons is an extremely slow process in this region. Hence the atomic ions undergo charge transfer reactions with neutrals as given below.

\[ A^+ + X_2 \rightarrow XA^+ + X \]  

(1.5)

Here \( A^+ \) is the atomic ions and \( X_2 \) denotes any neutral molecule. The rate coefficient (\( \gamma \)) of this reaction is of the order of \( 10^{-10} \text{ cm}^3 \text{ sec}^{-1} \). The molecular ion (\( XA^+ \)) thus formed undergo dissociative recombination with electrons as given by,

\[ XA^+ + e \rightarrow X + A \]  

(1.6)

The rate coefficient (\( \alpha \)), for the above reaction is of the order of \( 10^{-7} \text{ cm}^3 \text{ sec}^{-1} \).

At lower altitudes neutral density is more and hence equation 1.5 is faster and hence the overall rate is controlled by rate of equation 1.6, which is given as \( \alpha N^2 \), where \( N \) is the electron density. The layer thus formed is called \( \alpha \)-Chapman layer. At higher heights, reactions given in equation 1.5 become slow and overall reaction depends on the production rate of molecular ions. Here the rate of reaction is given as \( \gamma N N^+ \) or \( \beta N \) where \( \gamma N n = \beta \) and \( N n \) is the number density of neutrals. This layer is called \( \beta \)-Chapman layer. As we go high from lower altitudes, gradual transition occurs from \( \alpha \)-type to \( \beta \)-type process. At intermediate heights where \( \alpha \) and \( \beta \) type reactions are important, the production rate is given by

\[ \frac{1}{q} = \frac{1}{\beta N} + \frac{1}{\alpha N^2} \]  

(1.7)

The ionosphere is sometimes called a laboratory, where variety of such chemical reactions takes place, involving neutrals as well as ionized species. Some of the products of these processes may be in higher energy (excited) states, and, the subsequent de-excitation results in photon emissions, characteristic of the particular
interaction. These transitions could occur in the ultraviolet, visible or infrared regions of the electromagnetic spectrum, and are termed as *airglow*. Anders Angstrom was the first to note the weak radiations and discovered green line in the night sky in 1868. The detailed investigations of the spectral emission were done in 1920’s with the efforts of Lord Rayleigh. John McLennon and G. M. Shrum, in 1923, identified the green line to be originated from atomic oxygen. In the beginning of 1930’s, Sydney Chapman proposed chemical recombination as the mechanism for these lines. In 1950, the term airglow was coined to describe the emissions.

### 1.3 Airglow and Ionosphere

The airglow is generated through the photochemistry in the upper atmosphere, and the intensity of the emission depends on the strength of the reactants and the corresponding cross-sections. If the excitation process involves ionospheric plasma, then the airglow intensity can be represented as a function of the plasma (electron) density. Thus, the variations in the plasma distribution would reflect as intensity fluctuations of such emissions. This very fact is used or is made use of to determine ionospheric parameters from airglow measurements. It is known that 630.0 nm intensity is proportional to Ne and depends on height while 777.4 nm intensity is proportional to Ne^2.

Earlier works associating airglow with the ionosphere are from the observations of 630.0 nm enhancements, which were interpreted in terms of the descent of F-region. [Greenspan, 1966; Nelson and Cogger, 1971]. Higher F-layer altitude slowdown the dissociative recombination and results in less thermospheric emission [Bittencourt and Sahai, 1979]. Sharp gradients in 630.0 nm intensity results when F-layer height is modified by meridional circulation [Herrero and Meriwether, 1980]. Such studies point towards the relationship between the airglow emissions and the height of ionosphere, and show how these measurements give information about ionospheric and atmospheric dynamics. Coordinated experiments have been carried out to investigate thermospheric-ionospheric coupling by means of airglow, neutral winds, and ionospheric observations. For example, the study of brightness waves in airglow images and their relation to the
mid-night temperature maximum (MTM) [Colerico et al., 1996; Mendillo et al., 1997; Otsuka et al., 2003]. The association of MTM with atmospheric tides [Fesen et al., 1986; Fesen, 1996] throws light on the role of lower atmospheric process in controlling ionospheric behavior. Recent studies have shown the signatures of non-migrating tides in the global airglow distribution, thus revealing direct coupling between troposphere and ionosphere [Immel et al., 2006].

The interaction between neutral atmosphere and ionosphere sometimes also appear as periodic modulations of airglow, often associated large scale wave propagations known as traveling ionospheric disturbances (TID) [Garcia et al., 2000; Shiokawa et al., 2003]. TID’s are the perturbation of ionospheric plasma by acoustic gravity waves (AGW) [Hines, 1960]. The large scale AGWs or traveling atmospheric disturbances (TADs) are generated during geomagnetic storms, when the heating and subsequent rapid expansion of the high latitude results in pressure gradients and modify the thermospheric composition as well as circulation. These composition changes are transported to middle and low latitudes as TADs or AGWs [Roble et al., 1978; Prolss, 1987; Rishbeth, 1987]. Another common source of AGWs is in the lower atmosphere, often associated with weather fronts or land topography, which propagate upwards and sometimes perturb the ionosphere.

In addition to the modulations caused by external agencies, ionospheric plasma in the equatorial region is often re-structured by instabilities that operate when suitable conditions exist. These irregularities in plasma density appear as corresponding airglow patterns, and can be studied by measuring these intensity variations.

1.4 Ionospheric Irregularities
The nighttime equatorial ionosphere is characterized by irregularities of a wide range of scale sizes. After sunset, loss process in the ionosphere becomes predominant and sharp density gradients are generated. Also, the F layer is lifted up to very high altitudes due to the pre-reversal enhancement of zonal electric field. Under such conditions, instability processes can operate and render the plasma unstable. Any small
initial perturbations can grow in amplitude and will appear as ionospheric irregularities. These irregularities are collectively known as Equatorial Spread-F or ESF.

Rayleigh-Taylor instability is thought to be the mechanism responsible for ESF [Kelley, 1989]. The instability process could be understood with the help of Figure 1.2, which illustrates a simplified model of the post-sunset equatorial ionosphere, where the directions of density gradient, gravitational force and the magnetic field are indicated. This is a situation where the heavy plasma, on which gravity acts downward, is supported by the light magnetic field. The $g \times B$ force acting on the plasma results in an eastward current that depends on plasma density. Assuming sinusoidal initial density perturbation as given in Figure 1.2, the current will be more in the high density trough and less in the low density crest, causing the charge particles to be piled up at the walls of the perturbation. The polarization electric field thus generated will drive the low density plasma to higher altitudes and high density plasma to lower regions.

The irregularities are first detected from the spreading of ionogram traces [Booker and Wells, 1938], which explains the name ESF. This phenomenon occurs after sunset when F layer drifts upward to reach higher altitudes [Farely et al., 1970], and, manifest as ‘bite-outs’ in satellite density measurements [Hanson and Sanatani, 1973], plumes in radar RTI maps [Woodman and LaHoz, 1976], and cause scintillations of radio signals [Basu et al., 1978]. These irregularities convect upwards with large velocities through the F layer peak to the topside by flux tube interchange and can reach very high
altitudes [Woodman and LaHoz, 1976; McClure et al., 1977; Burke et al., 1979]. Tsunoda [1980] showed that the irregularities are field aligned and extend to hundreds of kilometers.

The ESF is associated with irregularities ranging a wide variety of scale sizes. The scale sizes vary from few centimeters to about a thousand kilometers. The irregularities are aligned along the magnetic field lines. They move upwards with velocities ranging from tens of $\text{ms}^{-1}$ to hundreds of $\text{ms}^{-1}$. Also they move eastward with velocity about 100 to 200 $\text{ms}^{-1}$. The small-scale irregularities associated with ESF cause scintillation in VHF signals. The large-scale structures are called plasma depletions. They represent a large-scale reduction in plasma density. Two to three orders of magnitude reduction in density are observed. They have vertical scale sizes of several hundreds to thousand kilometers. Their east-west scale size varies from tens of kilometers to hundred kilometers. They move upwards and eastwards.

1.5 Study of Plasma Depletions

Since the first observation by Booker and Wells in 1938, many investigators have looked into the phenomenon of ESF using various techniques. But, the mechanism responsible for the generation of irregularities over such a wide scale is still not completely understood. The occurrence of ESF is highly unpredictable. ESF occurs only on certain nights. The appearance of plasma depletions is even more unpredictable. Plasma depletions are not observed on all spread-F nights. This makes their study very important and interesting.

Ionosonde, scintillation receivers, radars, rocket and satellite borne probes etc. can be used to study the ionospheric irregularities. Ionosonde can be used to detect the onset of spread-F. But, once spread-F has set in, it cannot provide much information. Strong scintillations will be recorded in the scintillation receivers when the signals received traverse through these irregularities. The movement of irregularities can be studied using spatially separated receivers. But, this gives information about the smaller scale irregularities that give rise to scintillations. The plasma bubbles appear as plumes in the RTI maps obtained using VHF radars. Here also, the backscatter signals are due
to small-scale irregularities. Since the small-scale irregularities are found at the edges of large-scale structures, one can use the RTI maps to derive information about plasma depletions under certain assumptions. But field of view (FOV) is limited. Rocket and satellite borne probes are also used to study these phenomena. But they cannot give any information along perpendicular directions to their trajectory.

To get a complete picture about the formation of the depletions and their growth and movement an instrument that can cover a very large area of the sky is required. Optical imaging of the upper atmosphere is a powerful technique that can be used to understand this phenomenon completely. The technique is to take images of the night sky using the airglow emissions using an all sky imaging system. The imaging system employs a large field-of-view, front-end optics and hence is capable of covering thousands of kilometers of the sky and can be used to study the dynamics of plasma depletions.

1.6 Current Understanding

The signatures of ESF irregularities, which are the structures in plasma density distribution, should manifest as corresponding airglow intensity variations. Thus, ground based measurements of airglow intensity can be used to study these irregularities. Wide angle optical instruments were employed to image the auroral arcs at high latitudes [Mende and Eather, 1976], using television cameras. Similar technique was applied in the study of equatorial F-region using the 630.0 nm line. The observations revealed North-South aligned reduced airglow intensity, which are now called plasma depletions [Weber et al, 1978]. Further studies were conducted to understand the drift, strength (degree) and scale size of the depletions [Weber et al., 1980]. The depletions are considered as footprints of depleted flux tubes at the airglow emitting region. The reverse of this concept, the so-called ‘apex mapping’, was used to trace the poleward ends of depletions along the magnetic field lines to determine the corresponding altitude attained by the irregularities at the magnetic equator. This information was then used to determine the vertical rise velocity of depletions. The
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777.4 nm emission was also incorporated in the measurements for detailed investigations [Moore and Weber, 1981].

Ground based intensified all sky imagers were introduced in early 80’s [Mendillo and Baumgardner, 1982]. These observations revealed several features of depletions. The pronounced occurrence of depletions was observed between 2030 LT-2330 LT. The poleward end of depletion was found to have a westward tilt. Branching or bifurcation of depletions was also observed. Numerical simulations had shown that rising bubbles start bifurcating when contribution of E-region Pedersen conductivity to F-region conductivity is included in the calculations, and the bifurcations occur when the ratio of conductivities inside and outside the bubble becomes less than 10 [McDonald, 1981; Zalesak, 1982]. Detailed analysis showed that the conductivity ratio inside and outside the bifurcated depletion met the requirement, but the appearance of linear and bifurcated depletions side-by-side remained unexplained [Anderson and Mendillo, 1983]. The westward tilt of depletion was explained based on the latitudinal variation in zonal wind [Mendillo and Taylor, 1983]. Coordinated all sky imaging and scintillation observations showed that the irregularities producing scintillations are located on the walls of the bubbles [Weber, 1982].

With the establishing of ground based all sky imager, many investigators followed such observations, which revealed several characteristics such as sharp eastern walls [Rohrbaugh et al., 1989], day-to-day variability, high altitude (~ 2500 km) bubbles, post-midnight depletions associated with magnetic storms [Sahai et al., 1994], supersonic (2.8 km/s) vertical velocity, and absence of simultaneous 777.4 depletions with that in 630.0 nm [Weber et al 1996], etc. The association between spread-F in ionograms and irregularities in the walls of the depletions were demonstrated using ray tracing methods [Sales et al, 1996].

The trans-equatorial, meridional wind at the sunset terminator is believed to play an important role in the growth or suppression of the R-T instability [Maruyama, 1988]. When there is a strong N-S component, the ionospheric height decreases in the downwind side, hence the Pederson conductivity increases (more O₂⁺, NO⁺ ions), in comparison to the conductivity at the upwind side (O⁺ ions). This, results in an E-
region type of load at one end of the flux tube, and lowers the growth rate. The attempts to study the role of meridional wind in generating or inhibiting spread-F by simultaneous wind and depletion measurements did not show any conclusive evidence [Mendillo et al., 1992, Bittencourt et al., 1997]. It was found that during spread-F season, depletions were generated when large fluctuations were present in the meridional wind. During non-spread-F season, no depletion was seen when there were fluctuations in meridional wind. But, depletions were produced when magnetic storms were present; showing that when there is a strong seed present (magnetic storm) meridional wind also doesn’t suppress ESF. Several all sky observations have studied the occurrence of depletions and geomagnetic activity [Sahai et al. 1988; Aarons et al. 1999; Sinha et al., 2000, 2001].

The all sky measurements generally show eastward drifting depletions, but there are reports of deviation from this pattern when depletions reversed direction to westward for a period of about 1 hour [Taylor et al., 1997]. There are also reports of stationary depletions that do not show any movement for very long time period [Fagundes et al., 1999; Sinha et al. 2001]. The anomalies in the eastward drift are believed to be due to the modification of F-region dynamo by the changes in neutral circulation caused by magnetic activity. Sinha et al. [2003] reported splitting and joining of depletions, indicating strong shear in plasma drift. The zonal drift measurements are used to understand the latitudinal difference in plasma drift [Martinis et al., 2003; Pimenta et al., 2003a, 2003b]. The depletions were found drift at higher velocities near the equator compared to off-equatorial latitudes. The plasma drift measurements deduced from simultaneous 630.0 and 777.4 nm images showed slightly higher velocities near the F-peak altitude [Ablade et al., 2004].

The seasonal and solar cycle variations in the occurrence of depletions are only available from the Brazilian sector, owing to the extended data set recorded over that region. It is shown that the probability of occurrence of plasma depletions increases significantly during the periods of high solar activity [Sobral et al., 2002; Sahai et al., 2000].
1.7 Focus of the Current Study

The present work is aimed to study the various parameters of plasma depletions and to understand the role of various atmospheric and ionospheric conditions that result in the generation of these large-scale irregularities. The study is carried out using the all sky imaging system developed at the Physical Research Laboratory (PRL), Ahmedabad, India. The all sky imaging system of PRL was operated from Mt. Abu (24.5° N, 72.7° E; 18.5° N Magnetic) as well as Kavalur (12.5 N, 78.8 E, M.L. 4.6 N Magnetic), India. The observations were carried out during the increasing phase of solar activity in 1999-2002. The images obtained from this campaign were analysed to determine various parameters of plasma depletions such as degree of depletion, width of depletion, inter depletion distance and eastward drift velocity. Mt. Abu is an off-equatorial station whereas Kavalur is very close to the equator. Thus, the results give an opportunity to investigate the characteristics of depletions near the equatorial region as well as at off-equatorial latitudes. Over the equator, one could study the developing depletions or the initial stages of developed depletions. At off-equatorial location, the depletion would be already well developed.

The organization of the thesis is as follows. The mechanism of airglow emissions and all the technique of sky imaging are given in Chapter 2. The details of the PRL’s all sky imager is given in Chapter 3, which also gives an overview of the imaging observations used in the current study. Chapter 4 gives the characteristics of the plasma depletions from the observations conducted for this work, and Chapter 5 highlights some of the new observations and information revealed from these investigations. A summary of the results are given in Chapter 6.