CHAPTER - 1

INTRODUCTION

Much of the current research in the aeronomy of earth's atmosphere has been devoted to the study of the dynamics of the upper atmosphere. The objectives of this work are directed to the understanding and elucidation of how the various energy contributions to the earth's energy budget may be redistributed and how the composition of the atmosphere may be affected by the dynamical processes. The combination of extensive ground-based observational studies, recent opening up of the range and extent of experimental observations to the global scale brought about by the satellites and major advances in the development of theoretical models, have brought the field to a high degree of maturity.

Temperature is one of the basic parameters of the upper atmosphere. A study of its temporal variations and changes during specific ionospheric/geomagnetic conditions can provide significant insight into the energetics and dynamics of the upper atmosphere.

1.1 Temperature profile of Earth's Atmosphere

Based on the physical and dynamical properties at various altitudes, the earth's atmosphere has been
classified into distinct regions. The nomenclature generally used to designate these regions, which are layers with definite properties, is the direct outcome of the study of the temperature profile of the atmosphere. Figure 1.1 illustrates the typical variation of the atmospheric temperature with altitude. Also shown are the pressure, density and mean molecular weight in the atmosphere.

The lowest atmospheric layer is the troposphere in which the temperature falls off at a rate of $10^0 \text{K km}^{-1}$ or less and is bounded by the tropopause at a height of 10-12 km. Above tropopause is the region of stratosphere where there exists a temperature inversion with the temperatures increasing up to about 50 km. The stratopause at 50 km is due to the maximum absorption of solar UV radiation by ozone. Temperature decreases again in the mesosphere to the second minima at the mesopause (80-85 km), with temperatures around 180°K. This is the coldest part of the atmosphere. Above the mesopause, temperature gradient remains positive. This is the thermosphere, wherein the temperature rises sharply to about 1200°K at around 300 km and then becomes constant with altitude at this value which changes with time and other physical processes in the upper atmosphere. The thermosphere and mesosphere are the principal regions of interest for the present study since they play an important role in understanding the dynamics of upper atmosphere.
Fig. 1.1: Typical variation with altitude of atmospheric pressure $p$, density $\rho$, temperature $T$, and mean molecular mass $M$ up to 300 km. (Taken from H. Rishbeth and D. K. Garriott, Introduction to Ionospheric Physics, Academic Press, 1969).
1.2 Energy inputs in the Upper Atmosphere

1.2.1 Quiet times

The energy sources which have a dominant role in the dynamics of the upper atmosphere can be classified as follows:

(i) Absorption of solar ultra-violet radiation and X-radiation, leading to photo-ionisation, photo-dissociation and consequent chemical reactions, liberating heat.

(ii) Dissipation of tidal motions, gravity waves and release of energy from turbulent motions in the lower atmosphere.

(iii) Joule heating by ionospheric current systems, especially equatorial electrojet at low latitudes and auroral electrojet at high latitudes.

(iv) Absorption of energetic charged particles entering the atmosphere.

Solar Radiation

Solar radiation is the most important source of thermal energy in the upper atmosphere. The incoming radiation from the sun is absorbed by the atomic and molecular constituents of the atmosphere in different wavelength regions. Heating due to the absorption of solar EUV radiation is the dominant energy source for the thermosphere above about 150 km and heating due to photodissociation in the Schumann-Range continuum and absorption in the Schumann-Range bands are dominant below 150 km.
Previous calculation of the efficiency of solar EUV heating by Chandra and Sinha (1973) indicated an EUV heating efficiency of 50% while the studies of Stolarski et al. (1975) and Stolarski (1976) gave 30-40%. Recently, Torr et al. (1980a, b, 1981) reexamined the neutral heating efficiency due to solar EUV and UV absorption within the thermosphere. Their results showed that the heating efficiency is altitude-dependent, with a maximum value of 50-60% occurring near 200 km. The neutral gas heating efficiency is also time dependent, as discussed by Torr et al. (1981), because of the diurnal variation of minor neutral constituent densities and their rate of reaction during the day. Maximum heating efficiencies occur near evening twilight in response to the build-up of minor neutral constituent densities during the day. This shifts the time of maximum heating from noon toward the afternoon by about 2 hours.

Torr et al. (1980c) have presented measurements of solar flux in the Schumann-Runge continuum (1350-1750 Å) made by the Atmospheric Explorer satellites, which indicate a strong dependence on solar activity. Also Rottman et al. (1982) presented full-disk solar irradiance measurements obtained by the Solar Mesosphere Explorer (SME), indicating that the Lyman-alpha, Schumann-Runge continuum and the integrated Schumann-Runge bands (1750-1900 Å), all show variability related to the
27 day solar rotation period.

**Atmospheric Tides**

Lindzen (1967) had estimated that the flux of energy associated with diurnal tides is comparable to those of gravity waves. Lindzen and Blake (1970) have shown how the thermospheric temperature profile is modified by the tidal energy input in comparison to heating by solar UV radiation. The gravitational tides of sun and moon, but for their historical importance, are very minor sources of energy input in the upper atmosphere. Recently, Forbes (1982a,b) formulated a comprehensive tidal model that extends from the ground to 400 km and used it to investigate the solar and lunar diurnal/semidiurnal tidal components. Although there is a considerable day to day variability in the data, the model has good agreement with the observed 'mean diurnal tidal structure'. Reasonable agreement was obtained with various semidiurnal tidal modes and the results indicate that the semidiurnal tide has its origin in three sources of comparable importance: (i) solar EUV radiation in the lower thermosphere (100-200 km) (ii) in situ momentum coupling due to interaction of diurnal winds and diurnally varying ion drag, and (iii) tides propagating upwards from below 100 km.
Atmospheric Gravity Waves

At altitudes above 100 km, additional sources of heating, such as propagating gravity waves which dynamically carry the energy upward through the mesosphere, are available (Hines, 1960). It has been shown by Hines, that the heating due to the dissipation of these waves by molecular viscosity and thermal conductivity is sufficient to modify temperatures in E-region and to account for very rapid rise of temperature in the lower thermosphere. Hines (1965), using vertical wind profiles (Kochansky, 1964), estimated that the dissipative heating associated with gravity waves (which have periods of a few minutes to a few hours and spatial scales of a few km to some tens of km in the vertical direction and some hundreds of km in the horizontal direction), is about $10^0$K day$^{-1}$ at 95 km level, increasing to $100^0$K day$^{-1}$ near 140 km. Zimmermann and Rosenberg (1971) have calculated the rate of viscous damping from 70 mid-latitude wind profiles, which show large energy contribution near 95 km, tapering-off at higher altitude. They estimate this source to be four times greater than the Schumann-Runge continuum absorption in that region. Recently Richmond (1978a, b) made a comprehensive study of the properties of atmospheric gravity waves, using both theoretical analyses and numerical simulations.
In the first paper (1978a) he examined the nature of gravity waves ducting in the thermosphere and in (1978b), examined the generation, propagation, and dissipation of long period (> 30 min.) gravity waves, using an analytical approach. Walterscheid (1981) found that vertically propagating internal gravity waves in the upper atmosphere induce a downward transfer of significant heat from regions of wave dissipation and suggested that this transfer of heat may result in a net cooling of regions of the upper atmosphere. However, his analysis considered temperature changes induced by a single wave (Roble, 1983). A broad spectrum of waves with different phase velocities may produce substantial cancellation. Atmospheric gravity waves may produce a net heating of the upper atmosphere due to viscous dissipation. However, the significant heat transport by the wave may mitigate the heating and for certain characteristic wave quantities may even produce a net cooling in the regions of the lower thermosphere. Such dynamic processes could account for the inverse correlation of lower thermospheric temperature with solar F10.7 flux values as observed occasionally by Hernández, 1976 and Chanin and Tulinov, 1979. This is possible if the gravity wave energy is somehow related to the solar F10.7 flux (Roble, 1983).
Joule Heating and Particle Precipitation

In the dynamo-region of the ionosphere, the currents that are driven by tidal winds, produce joule heating. Cole (1962 a,b) estimated the dissipation due to these currents and found it to be comparable with the Hines (1965) estimate of semi diurnal tidal flux. The effect of joule heating for exciting tidal modes and gravity waves has been investigated by Blumen and Hendl (1969), and Sastry et al. (1973) have explained in part their observations of abnormal electron temperature measurements at equator as due to joule heating. The equatorial electrojet as a source of gravity waves has also been proposed by Chimonas (1970).

Energetic particles, right from auroral to cosmic ray energies, constantly bombard the upper atmosphere. This phenomenon is most pronounced at high and auroral latitude. Energy deposition associated with these energetic particles in an active aurora can exceed, at times, the EUV flux by a factor of $\sim 100$.

1.2.2 Geomagnetically disturbed times

Large currents flow in the auroral zones at times of geomagnetic disturbances and the joule heating from such currents had been estimated to be few tens of ergs cm$^{-2}$s$^{-1}$ (Kennel and Rees, 1972). Large amounts of energy, both particle and field contributions, are deposited in the lower thermosphere and cause large scale dynamic
motions that result in the overall global heating associated with magnetic storms (Hays and Roble, 1971; Hays et al., 1973). At times of magnetic storms and active auroral displays, it has been shown (Chimonas and Hines, 1970) that energy might be transported from the auroral zones to lower latitudes by gravity waves and thereby provide excess heating. Richmond (1979b), recently carried out numerical simulations concerning the transfer of energy from high to low latitudes during geomagnetic storms. His results show that dissipative effects progressively remove more and more of the slower moving and shorter-period waves generated by the source and that the long-period waves with high horizontal velocities are able to propagate farthest in latitude in the thermosphere before being dissipated. Also during geomagnetic storms, meridional circulation cells are generated (Rishbeth, 1974) and these can effectively transfer the energy inputs at higher latitudes to lower latitudes. A detailed discussion on this special topic of energy transport from high to low latitudes during geomagnetically disturbed times, will be given in the next chapter, wherein our observations on excess equatorial thermospheric heating on such occasions is dealt.

Torr and Torr (1979) suggested that fast O\textsuperscript{+} bombardment of the thermosphere is a significant heat
source during geomagnetic storms and the thermospheric response to the heat source was discussed by Torr et al (1982). Tinsley (1979a,b, 1980, 1981) has pointed out that neutral hydrogen atom precipitation due to magnetospheric ring-current decay may be a significant heat source for both the low and mid-latitude thermosphere. The Dynamic Explorer satellite, which measures the neutral gas composition (Carignan et al., 1981), has shown a strong magnetic control of the composition in the high latitude thermosphere with a universal time dependence that extends to mid-latitudes (Carignan et al., 1982).

1.3 Atmospheric Models

Several models of earth's atmosphere have been developed, which represent the temperature, density, pressure, concentrations etc. as a function of time, altitude and latitude. As a first step towards atmospheric modelling, it was necessary to have a semi-empirical representation of the above parameters based on relatively simplified physical concepts. One of the results of such attempts is to permit the introduction, as known, of the parameters in the partial solution of the physical equations.

This was the basis of the first models, notably those of Nicolet (1961, 1963). These models suppose that temperature and composition are fixed at 120 km (the most reasonable hypothesis possible at that time).
Above this altitude, the number densities of individual species follow the theory of molecular diffusion (i.e. diffusive equilibrium) except for hydrogen, where equilibrium is obtained at higher altitudes (Kockarts and Nicolet, 1962, 1963). Thermal diffusion is taken into account for helium. The work of Harris and Priester (1962), in which more physical concepts were introduced, is also worth mentioning. The hydrostatic equation and the heat conduction equation are integrated simultaneously with a varying heat source of a 24 hour cycle. However, it is necessary to adopt a second heat source to represent the densities correctly; this was the base of the international reference model (CIRA, 1965). For this reason, in 1965, Jacchia preferred to abandon the theory and return to a more simplified physical concept. He constructed an empirical temperature profile to represent the measured values, mainly total densities, taking Nicolet's models as a pattern. Such a semi-empirical model already gave a satisfactory representation of the variations of the parameters with the correct order of magnitude. Systematic errors existing in the first models as Jacchia, 1965 (J65) were eliminated in the later models by virtue of a large amount of ground based or in-situ satellite data deduced from mass spectrometer, accelerometer, optical techniques, and incoherent scatter radar measurements. It became clear
that any progress in the empirical representation required the lower boundary conditions to vary at 120 km and that it was necessary to model these variations or to assume conditions fixed at lower altitude (90 km, Jacchia, 1971). Nowadays, in most cases, the temperature and concentration variations at 120 km are represented by expansions in terms of spherical harmonics, as proposed by Hedin et al. (1974). Some of the semi-empirical models will be briefly described in the following paragraphs.

The present empirical models of the thermospheric temperatures and densities are based on artificial satellite drag (Jacchia 1970, 1971, 1977; CIRA 1972; Barlier et al. 1978), mass spectrometer density (Hedin et al. 1974, 1977, 1979; Von Zahn et al. 1977; Kohnlein, 1980), optical measurements of Doppler width of atmospheric emissions from an artificial satellite (Thuillier et al. 1977a), incoherent backscatter radar measurements (Oliver, 1979), as well as combinations of some of the above (Hedin et al. 1977, 1979; Thuillier et al., 1977b; Barlier et al., 1978). Of these, only the model by Thuillier et al. (1977a) directly models neutral kinetic temperature (i.e. Doppler widths of line emissions) as a measured quantity. The other models find temperature as a derived quantity, such as temperature derived from thermal plasma fluctuations, the scale height temperature of an atmospheric constituent, or that 'temperature'
which gives agreement with the drag derived densities, and in the last two cases the assumption of a static atmosphere in diffusive equilibrium is made.

As would be expected, these models vary considerably among themselves in the description of thermospheric temperature and its behaviour, which has led to critical examination of these models, such as surveys of Von Zahn and Fricke (1978), Jacchia (1979), Barlier et al. (1979), Hickman et al. (1979) and Barlier and Berger (1983).


The interaction of the thermosphere with particle precipitation, joule heating, etc. are incorporated in the models by inclusion of any of the indices of magnetic activity $K_p$, $a_p$ and $A_p$ as parameters.

Parallel to the development of the semi-empirical models, a considerable effort has been made to calculate theoretical models by solving the physical equations which govern the thermosphere (vide: Fuller-Rowell and Rees 1980; Fontanari, 1981 and Roble et al., 1982). These models have achieved the goal of a three-dimensional time dependant representation of thermospheric dynamics.
To conclude this section on atmospheric models, it is worth mentioning that the two most popular models commonly used to compare ground based and satellite measurements of the thermospheric temperature are the mass spectrometer and incoherent scatter MSIS (Hedin et al., 1977) and J77 (Jacchia, 1977). Hernández (1982b) had carried out an extensive comparison of thermospheric kinetic neutral temperatures measured by ground based Fabry-Perot spectrometer on OI emission. He concludes that of the available models (including MSIS and J77), in regard to temperature, the J77 model best represents the data set although the model is based on least detailed information. Keysor et al. (1979) from the Atmosphere Explorer (AE) satellite data have also found that temperatures in the altitude region of 150 to 225 km are best represented by J71 (Jacchia, 1971, which is an earlier version of J77) model, than does the prediction for this region by the MSIS model.

1.4 Determination of Upper Atmospheric temperatures

There are two fundamental approaches to the determination of upper atmospheric temperatures. In one case the 'in-situ' temperature is determined by observations of the response of the atmosphere to the insertion of a probe placed within the atmosphere. The probe could be of the direct nature, such as a mass spectrometer sensor or a released chemical. Also it could be of the indirect type as inference of the atmospheric temperature from satellite drag measurements.
The other approach is the ground based measurement wherein the observation of the electromagnetic radiation emitted or scattered by atmospheric species could be monitored by ground based instrumentation. Here an example of the direct type of measurement would be the Doppler shift and width of 6300 Å emission from F-region heights while obtaining ion temperatures from incoherent scatter radar data would be of the indirect type.

Some of the in-situ and ground based techniques will be illustrated in the following paragraphs with emphasis on the ground based spectroscopic and in-situ chemical release methods for upper atmospheric temperature measurement. These two methods were utilized in the present investigations.

1.4.1 Ground based methods

The two major instruments developed in the last two decades that have been highly successful in ground-based studies of the upper atmosphere by means of remote sensing are the incoherent scatter radar and the Fabry-Perot interferometer. Measurement of atmospheric motion with these instruments are based upon observations of the Doppler shift of electromagnetic radiation emitted by the medium as a result of natural processes or scattered by the medium from a signal supplied by the observer i.e. a radar.
Incoherent scatter

Conventional ionospheric sounding depends on the reflection of radio waves by an ionized gas and this reflection process depends on the collective behaviour of electrons, which can be approximately described in terms of the refractive indices of the Appleton-Hartee equation. A second kind of echo, much weaker than the first, arises due to the scattering from irregularities or sharp gradients of ionization when the radio waves frequency exceeds the local plasma frequency. A third kind of echo, even weaker than the others, is due to the Thomson scattering of waves from individual electrons. This is also called the incoherent scatter. The radar instrument observes the Doppler Shift in radiation received by the antenna after the original transmission from the antenna as a pulse and incoherently scattered within the ionosphere by small scale density fluctuations of electrons (Evans, 1974). Since the time over which the return echo of the scattered radiation is observed can be controlled by the radar receiver, information on electron densities, ion and electron temperatures, and ion drifts may be obtained as a function of altitude. The incoherent scatter return spectrum contains all this information. From the width of the spectrum, both ion
temperature and mass can be obtained. If the nature of the dominant ion is known, its temperature $T_i$ can be determined. Since the ions and electrons in the ionosphere are not usually at the same temperature, $(T_e \neq T_i)$, the scatter spectrum has two maxima. The theory of this instrument has been extensively developed and details can be found in the reviews by Evans (1974, 1970) and Walker (1979).

The first major incoherent scatter facility was constructed at Jicamarca, Peru in early 60's. Later, many more facilities were developed particularly at high latitudes viz. St. Santin, Millstone Hill, Chatanika and EISCAT. At low latitude, another important station is at Arecibo ($10^\circ$N).

Roble et al. (1970) have studied the ionospheric effects of the gravity wave launched by a sudden commencement of a geomagnetic storm, by using ion and electron temperature measurements from the Millstone Hill and Arecibo radars. Recently, Mazaudier and Bernard (1985), have reported incoherent scatter measurements of ion drift velocities obtained by St. Santin-Nancay radar during geomagnetic storms. A storm time meridional circulation with equatorward wind velocities at higher altitudes ($\geq 135$ km) is revealed in their measurements.
Hernandez et al. (1975) and Wickwar et al. (1984) have compared the neutral temperatures and winds measured by the incoherent scatter and Fabry-Perot techniques. The results show high degree of correlation with a difference of only about 30°K between the two types of measurements.

**Spectroscopic Technique**

Unlike the incoherent scatter radar, the Fabry-Perot spectrometer uses the light emitted by natural processes within the atmosphere to determine the line of sight Doppler shift and widths. The source of this light is the airglow emissions contributed by atmospheric chemical processes. For the F-region study, neutral oxygen line at 6300 Å is convenient and most of the ground based high-resolution work has been done on this emission by using Fabry-Perot spectrometers.

One of the first measurements of upper atmospheric temperatures from doppler line widths using a photo-electric Fabry-Perot spectrometer are those of Turgeon and Shepherd (1962) and Jarrett et al. (1964). Following these pioneering efforts, Karandikar (1968), and Biondi and Feibelman (1968) from the observed line profiles obtained no evidences for a dissociative origin of the \(^1D\) or \(^1S\) oxygen atoms. Also there was a possible indication of F-layer wind velocities of \(\sim 400 \text{ ms}^{-1}\).
Armstrong (1969) measured eastward winds in the evening, equatorward winds from midnight to near dawn and westward winds just before dawn. On a number of nights, the maximum measured wind speeds were in harmony with models of Kohl and King (1967), and Challinor (1968).

Hays and Roble (1971) had made direct observations of thermospheric winds during geomagnetic storms. One fringe profile was obtained in the zenith where the component of the wind along the line of sight of the instrument is small. The other fringe profile was obtained at a high zenith angle where a component of the horizontal wind exists along the line of sight of the instrument.

Rajaraman et al. (1978) described the preliminary results of F-region nighttime temperatures obtained with a pressure scanned Fabry-Perot spectrometer operated at tropical latitudes (Mt. Abu). Their results showed very good correlation between occurrence of spread-F over an equatorial station and increase in temperature over Mt. Abu.

Hernandez and Roble (1976) made measurements of thermospheric temperature and winds during geomagnetically quiet and stormy periods. These were compared with three dimensional semiempirical model of the neutral thermosphere. During geomagnetic storms, the nighttime equatorward winds are generally enhanced from their quiet time values with a maximum measured velocity of
$\sim 600 \text{ ms}^{-1}$ during a $K_p=9$ storm. The neutral temperatures also show enhancement from the quiet time values. Sipler and Biondi (1979) have reported similar enhancements in equatorward winds and neutral temperatures.

Sipler et al. (1982, 1983) have studied the mid and equatorial thermospheric temperature and wind fields over an extended period during 1977-79. A comparison with the NCAR thermospheric general circulation model (TGCM) of Dickinson et al. (1981) show reasonable agreement with the measured temperatures and winds.

One of the most exciting breakthroughs in optical interferometry was made by Hernandez et al. (1981) and Hernandez (1982d) that led to the development of the twin etalon scanning spectrometer (TESS). The light throughput in TESS is increased over that obtained by a single-etalon Fabry-Perot interferometer by a factor of $\sim 20-30$. As a result, the temporal resolution for wind and temperature measurements is greatly improved. Using the TESS instrument, Hernandez (1982d) detected oscillations in the vertical winds with a periodicity of about 40 minutes and with a magnitude of 50 ms$^{-1}$. These motions are ascribed to the passage of gravity waves over the observational site (Fritz Peak Observatory) during quiet geomagnetic conditions.
The restriction of the Fabry-Perot interferometer to nighttime observations can be removed by the use of multiple etalon interferometry (Cocks and Jacka, 1979; Cocks et al. 1980; Hays, 1982; Rees et al. 1982), and daytime neutral wind observations from ground have been reported by Cocks et al. (1980).

Rees et al. (1984) have developed a Doppler Imaging System during 1982 and made observations of auroral thermosphere with this new type of instrumentation. As claimed by the authors, this technique provides a multiplex gain of about 20, larger than can be obtained using a mask to scan multiple fringes in a Fabry-Perot image plane (Okano et al., 1980; Biondi et al., 1985 and Sahu et al., 1985) and at least 100 times that of the classical interferometer using central aperture only. Of course, this could be debatable as outer zones being extremely narrow, would contribute more to the noise and thereby making this technique comparable to the masking technique.

Some of the very recent work on upper atmospheric dynamics using Fabry-Perot techniques are reported in Cogger et al. (1985); Yagi and Dyson (1985a,b); Biondi and Sipler (1985); and Okano et al. (1985).

Temperature measurements in the mesosphere by spectroscopic technique of studying the band spectra of emitting species are reviewed by Meriwether (1984).
This technique along with our own measurements are reported in Chapter 4.

1.4.2 In-situ methods

In-situ measurements for temperature and winds (direct or indirect) of the upper atmosphere have been carried out by two means. The first method was suggested by Bates (1950) and has been used frequently from the time of the early rocket launches. A chemical compound is released into the atmosphere from a rocket vehicle and allowed to react with the ambient molecules and generate luminous trails or clouds. Ground based photography against the background field of stars over a period of time will establish a track signifying the direction and speed of the emission source.

The other in-situ technique consists of using a neutral mass spectrometer on board a satellite. From the measured neutral number densities and considering the diffusive equilibrium, neutral temperatures can be obtained. Atmospheric temperatures can also be obtained by the considering satellite drag and subsequent orbital decay. Further, with the introduction of satellite based Fabry-Perot instrumentation, direct measurements of thermospheric winds and temperatures are now possible.
Chemical Release

Heppner and Miller (1982) have reported a compilation of 39 chemical release experiments for thermospheric winds at high latitudes. Desai and Narayanan (1970) have reported low latitude (Thumba, India) diffusion and temperature measurements using Ba-Sr cloud releases. Further details on this technique and our own results on releases of Ba-Sr and Na are discussed in the next chapter.

Satellite based method

The advent of the Dynamics Explorer satellites, has provided a platform by means of which the dynamics of the ionosphere and the thermosphere can be studied with global coverage. The combination of the two instruments, the Fabry-Perot interferometer (Hays et al., 1981) and winds and temperatures spectrometer (WATS) (Spencer et al., 1981) on these satellites yields information on meridional, zonal and vertical components of thermospheric wind.

With the observations made from space platforms, recently Killeen (1985) has reported the strong dependence of thermospheric dynamics on forcing processes of magnetospheric origin.

1.5 Scope of the present study

In the present study, the following investigations are carried out:
(i) Measurement of diffusion coefficients and temperatures of the equatorial thermosphere by point release of Ba-Sr clouds.

(ii) Diffusion and temperatures observed from a Na trail release on an occasion two hours after a storm sudden commencement (SSC).

This particular experiment was fortuitous as both releases were planned for pre-ESF (Equatorial Spread-F) conditions.

Excessive equatorial thermospheric temperatures are observed on the Na trail release day. This cannot be explained by considering the gravity waves or meridional circulation mechanisms for energy transport from high to low latitudes during geomagnetic storms. An alternative mechanism, involving the hydromagnetic heating of thermosphere associated with sudden commencement is proposed, which satisfactorily explains the excess temperatures.

(iii) Nighttime F-region winds and temperatures, by monitoring the Doppler width and shift of OI 6300 Å airglow line, using a Fabry-Perot spectrometer.

(iv) Study of mesospheric airglow emissions, revealing the excitation mechanism for hydroxyl bands and indication of gravity wave passage through the mesosphere.