Chapter 5

Measured and derived meridional winds and their variabilities

5.1 Introduction

In his paper to Nature, Jacchia [1959] first pointed out the possibility of a difference in density of the upper atmosphere between the bright and dark hemispheres of the earth, which had led to the orbital decay of artificial satellites. It was recognized by then that the variations in air density are primarily due to heating by solar EUV radiation. The existence of a diurnal bulge of atmospheric density above about 200 km was attributed to the thermal expansion of the atmosphere at heights above 100 km in response to the large diurnal variation of temperature in this region. This 'bulging' of the atmosphere was shown to be responsible for the horizontal gradients of air pressure which would then drive horizontal winds [King and Kohl, 1965; Geisler, 1966]. Kohl and King [1967] showed the existence of a global wind system at F region heights through calculations using data from Jacchia's model atmosphere. They demonstrated the importance of ion-drag as a force determining the form of the atmospheric wind system.
Rishbeth [1972] in his review, examined the various properties of the wind system and the effects of the winds on the ionospheric $F_2$ layer. Experimental and theoretical results have suggested that the large scale dynamics arising out of such wind systems, would in turn, influence the temperature and composition of the thermosphere [Shimazaki, 1972; Mayr et al., 1978].

Apart from the role of meridional winds in moving the ionization along the magnetic field lines thereby altering the plasma density distribution, the neutral wind has many other effects on the equatorial and low latitude $F$ region. The winds are responsible for the global dynamo action leading to the generation of electric fields. The thermospheric zonal winds act as a primary driving force for the $F$ region dynamo, which gives rise to the enhancement of the post-sunset equatorial electric field [Rishbeth, 1971a,b; Heelis et al., 1974]. A transequatorial wind would transport ionization from one hemisphere to the other, thereby making the distribution of ionization asymmetrical about the equator [Bramley and Young, 1968]. It thereby aids the renewed fountain action in one of the hemispheres in the evening hours and enhances the crest of ionization. This feature has been shown to result in increase in peak electron density at a low latitude station, Tahiti, [Buonsanto, 1988]. Another effect of the transequatorial wind is to suppress the generation of ESF irregularities by enhancing the $E$ region Pedersen conductivity in one of the hemispheres and shortcircuiting the polarization electric field [Mendillo et al., 1992]. Recently, it has been shown that the vertical winds at the equator have significant role in enabling/inhibiting the Rayleigh-Taylor instability which is responsible for the generation of irregularities in electron and ion densities (Equatorial Spread $F$ (ESF)) [Raghavarao et al., 1987, 1992; Sekar and Raghavarao, 1987].

An outcome of the interaction of the global scale wind with anomalies in ionization distribution is a circulation associated with the temperature and pressure
perturbation [Dickinson et al., 1971]. The horizontal convergence (divergence) of the neutral wind forces a downward (upward) vertical wind in the immediate region of the ionization anomaly. Such motions would lead to adiabatic heating and cooling within the location of anomaly. Recently, Raghavarao et al. [1991] have reported an Equatorial Temperature and Wind Anomaly (ETWA) at thermospheric altitudes characterized by the formation of two regions of enhanced temperature on either side of a prominent trough over the magnetic equator. The zonal winds were shown to reach their maximum near the magnetic equator with two minima on either side of it. The trough in temperature and the maximum in zonal winds were seen to be collocated with the trough of the equatorial ionization anomaly. In a more recent paper, Raghavarao et al. [1993] proposed two whorls of meridional winds at each of the temperature (and pressure) crests. These winds are superposed on the planetary scale meridional winds due to diurnal tide. Thus the importance of measuring neutral winds and their variation with time in the low latitude region can hardly be over-emphasized.

In the following section (5.2), the various techniques available for wind measurements are discussed. In section (5.3), a brief review on the results obtained by several workers from ground-based optical studies for low and equatorial latitudes is given. This is followed by a discussion on a few indirect methods available to infer neutral winds (section (5.4)). For the low latitude stations, Mt. Abu/Ahmedabad, under the present study, we utilize direct measurements of meridional winds with the Fabry Perot Spectrometer, and the meridional winds derived using ionospheric data, for a detailed study on their (meridional winds) behaviour and their interaction with the ionospheric F region. The results are presented in sections (5.6) and (5.7). The method adopted for deriving meridional winds from the ionospheric data is described in section (5.5). The variation in neutral temperature is shown to be important in the procedure of deriving meridional winds, more so when specific events are studied.
(section (5.6)). However, for gross, average, features, use of the model temperature itself is adequate. This conclusion has been arrived at from a critical study to be discussed here (section (5.7)). The seasonal dependence of the magnitude and direction of meridional winds (both inferred and directly measured) is presented in section (5.7).

5.2 Sources of experimental data on neutral winds

One of the oldest methods to measure neutral winds in the upper atmosphere is by releasing vapour clouds from a rocket, as was first suggested by Bates [1950] and had thus proved to be the most fruitful method at height range of 120–200 km. The vapours most often used were sodium and trimethyl aluminium. Few of the limitations of the vapour trail method are that observations from at least two locations are required in order to determine, by triangulation, the position of an identifiable point in the release, and releases like that of sodium are confined to twilight periods only. It further requires cloud-free skies over a large area during the experiment. In spite of these stringent requirements, many series of observations have been successfully completed and yielded good results [Jarrett et al., 1963; Bhavsar et al., 1969; Bedinger, 1972; Desai et al., 1975; Raghavarao et al., 1984, 1987, to state a few].

With the advent of satellites, a series of experiments was conducted to obtain in situ measurements of neutral atmospheric parameters using omegatron instruments on San Marco 3 and 4 and quadrupole mass spectrometers on Aeros A and B, Atmosphere Explorer C, D and E and Dynamics Explorer 2 [Spencer et al., 1981 and references cited therein]. In the later version of quadrupole mass spectrometers, measurements of the neutral winds were made through interpretation of the modulation of the particle stream entering the mass spectrometer, using the baffle technique. The accuracy of measuring neutral winds by this technique, i.e., with the Wind and
Temperature Spectrometer (WATS) on board the DE 2 satellite was estimated to be 10–20 m/s at 650 km altitude. Subsequently, Raghavarao et al. [1991, 1993] have shown that the error is only 2 m/s at 300 km as it is inversely proportional to the square root of the gas density. The scientific results arrived at and important contributions made by the DE 2 mission have been reviewed by Killeen and Roble [1988]. The measurement concept of obtaining the neutral atmospheric parameters using mass spectrometers has been discussed by Spencer and Carignan [1988].

Line profile measurements of natural airglow emissions using Fabry Perot Spectrometer (FPS) have been yielding thermospheric temperatures and winds for more than four decades, starting from the pioneering work by Armstrong [1953]. The Fabry Perot interferometers have also been flown on satellites such as Dynamics Explorer 2 and have yielded good data [Killeen and Roble, 1988]. In their review article, Hernandez and Killeen [1988] have given the historical development of optical remote sensing with FPS as a tool and discussed the scientific progress made from the data base collected at various observing stations round the globe. As pointed out by them, although the Doppler measurements of winds and temperatures provide continuously a synoptic picture of dynamic and thermodynamic behaviour of the upper atmosphere, they (the measurements) assume utmost importance when interpreted in conjunction with a global circulation model.

5.3 Results from previous ground-based optical studies from low and equatorial latitudes

The ground-based measurements from the near equatorial zone are not many, when compared to those in high and midlatitudes. The only measurements reported in the literature in recent times are enumerated below. The results obtained by them are briefly reviewed.
Sipler and Biondi [1978] and Sipler et al. [1983] have made measurements over a span of several years at sites such as Kwajalein Atoll in the Marshall Islands (9.4°N, 167.5°E, geographic; 8.6°N dip latitude). They showed that the winds in the equatorial region are strongly influenced by variations of ion-drag due to the variations in the ion densities effected by $E \times B$ drifts, upward propagating tides from the lower atmosphere and also by high latitude magnetospheric convection. They have made this suggestion after comparing their results with the NCAR thermospheric general circulation model (TGCM) predictions.

Some useful results have been obtained on the equatorial thermospheric dynamics by measurements carried out by Biondi and Sipler [1985] as a part of the Brazil Ionospheric Modification Experiment (BIME) from Natal (5.9°S, 35.2°W, geographic; 6.4°S dip latitude), Brazil. They observed new effects such as (a) significant vertical flows and neutral temperature increases in response to persistent convergence in horizontal flows and (b) gravity wave modulation of thermospheric flow. These results, although of a single night measurements, are considered to be extremely important.

Meriwether et al. [1986] have obtained nighttime measurements of horizontal winds at Arequipa (16.5°S, 71.4°W, geographic; 3.5°S dip latitude), Peru, on 62 nights. Comparison with the predictions of NCAR TGCM model for equinoctial and solstice conditions showed good agreement. In their earlier study on O I 6300 Å nightglow brightness at Arequipa [Meriwether et al., 1985], they have examined the effect of a prolonged equatorward wind (~ 100 m/s) on the equatorial ionosphere. They have observed widespread areas of airglow depletion, with reduction in intensity as large as factors of 3 or 4. Ruling out the possibility of association of these depletions with equatorial plasma depletion events, they have invoked the presence of an equatorward neutral wind which would drive the $F$ region plasma up the magnetic field lines sufficiently to account for the intensity depletions.
As part of the CEDAR (Coupling, Energetics and Dynamics of Atmospheric Regions) program, Biondi et al. [1988] have carried out coordinated measurements of the dynamical behaviour of the upper atmosphere and the ionosphere at the magnetic equator, using a Fabry Perot interferometer from Arequipa and the incoherent scatter radar from Jicamarca. Motions between the zonal neutral wind and the zonal drift velocity of the $F$ region plasma were compared. The operation of $F$ region dynamo in conjunction with the decrease in $E$ region Pedersen conductivity was invoked to account for the correlated motions of the neutral and plasma in the zonal direction during late night hours. During daytime, however, the $F$ region dynamo field is shorted by the higher conductivity in $E$ region accounting for the uncorrelated motions.

Continuous nighttime FPI measurements on the 6300 Å emission line from Arequipa have yielded a data base covering 2/3 of a solar cycle. Biondi et al. [1990, 1991] have reported monthly average variations in the meridional and zonal components of neutral wind, displaying seasonal changes in wind patterns. The measured seasonal variations in the wind patterns are more pronounced than the solar cycle variations, which are expected in terms of the underlying forcing and damping processes. At the winter solstice, they have observed a weak (~ 100 m/s) transequatorial flow in the earlier and late part of the night, with essentially zero velocities in between. At equinoxes, there is an early night poleward flow during solar minimum, which becomes equatorward during solar maximum. The zonal winds were observed to be predominantly eastward throughout the night, except during solar minimum equinoxes when there is a brief westward flow in the early and late part of the night.

Biondi et al. [1990] have compared their results with the horizontal wind model (HWM-87) of Hedin et al. [1988] and the vector spherical harmonic model (VSH) of Killeen et al. [1987] and found a general agreement. Comparison has also been
made with the winds measured at Arecibo (18°N, 67°W, geographic; 30°N dip lati-
tude) [Burnside and Tepley, 1989], the geographic ‘mirror twin’ of Arequipa. With
the solar EUV source acting as the principal driving force, the expected oppositely
directed meridional flows and similar eastward zonal flows at the two locations have
been reported by them.

Contradicting the conclusion arrived at by Burnside and Tepley [1989] that
nocturnal and seasonal variations in the neutral wind field are remarkably unaffected
by changes in the solar cycle, Biondi et al. [1991] demonstrated significant differences
in both meridional and zonal wind patterns at different seasons as the solar cycle
progressed. As pointed out by them, this is expected since the increase in pressure
gradients from solar minimum to solar maximum would produce higher wind speeds
near the subsolar point at locations such as Arequipa. However, the increase in the
zonal wind has been observed to be moderate suggesting the opposing role played by
ion-drag which would reduce the effect of the stronger pressure gradients. It is to be
noted that the meridional wind speeds at Arequipa are small throughout the solar
cycle, the wind flows near the subsolar point being dominantly in the zonal direction.
Also at times during magnetic disturbances arising out from the auroral regions, the
meridional flows would tend to cancel near the magnetic equator.

Though the ionosphere over Arecibo in Puerto Rico has midlatitude character-
istics (dip angle $I = 51^\circ$), the thermospheric dynamics in this region exhibits features
characteristic of low latitude behaviour. Optical observations at Arecibo have yielded
very useful results supporting the conclusions arrived at by satellite measurements.
Burnside and his coworkers [Burnside et al., 1981; Burnside et al., 1983; Burnside,
1984; Burnside and Tepley, 1989] have been reporting optical observations of night-
time thermospheric winds carried out from Arecibo since 1980. Apart from their
conclusion that the thermospheric wind fields at Arecibo do not appear to show any
solar cycle dependence, they have reported nighttime wind patterns controlled by the disturbance associated with midnight pressure bulge. The zonal winds are eastward throughout the night in winter. A reversal to westward flow is usually observed after local midnight in summer months. The meridional flows are largest in summer, while a reduction or sometimes reversal occurs after midnight. The data from the incoherent scatter radar have been used along with optical observations to infer the occurrence of midnight pressure bulge [Burnside et al., 1983]. A feature which is associated with the reversal of equatorward wind near midnight, is the meridional intensity gradient (MIG) in 6300 Å emission [Herrero and Meriwether, 1980; Friedman and Herrero, 1982]. The interaction of the disturbance originating in polar region and moving towards the equator, with the poleward winds associated with the midnight pressure bulge, leads to a transition region where large horizontal gradients in wind velocity and in the airglow intensity have been observed.

Sahai et al. [1992a] have reported thermospheric neutral wind measurements by observing O I 6300 Å nightglow emission at Brazilian locations, Sao Jose dos Campos (23.2°S, 45.9°W, geographic) and Cachoeira Paulista (22.7°S, 45°W, geographic) during 1988–1989. The dip latitude was about 16°S for these stations. These measurements turned out to be the first ever reported from a location near the crest of the equatorial ionization anomaly. Sahai et al. have displayed graphically the average nocturnal variations of both meridional and zonal winds during all seasons and compared their results with HWM-87 and HWM-90 models. On comparison, they have reported some discrepancies both in the absolute magnitudes and in details of nocturnal variations. The overall wind patterns were similar to those reported from Arequipa. Fig. 5.1 shows their results on seasonal variabilities of meridional winds over Cachoeira Paulista and the comparison with the models.
Fig. 5.1. Temporal variation of the measured meridional winds for different seasons over the Brazilian latitudes (23°S, 45° geographic; 16°S dip latitude) (after Sahai et al., 1992a).
5.4 Indirect methods available to infer neutral winds

Apart from the direct ground-based optical techniques, several other methods are available to determine neutral winds; from (i) ion velocity measurements of incoherent scatter radars, (ii) satellite measurements of O I 6300 Å emission, and (iii) F layer height and its variation deduced by ground-based ionograms. There are several assumptions and limitations in each of these methods. With more data on direct measurements of winds being accumulated, these methods are refined and they help in further advancement of our knowledge on the thermospheric dynamics. These methods are briefly reviewed below.

(i) Neutral winds deduced from incoherent scatter radar (ISR) observations

The method of employing F region ion drift measurements obtained by incoherent scatter radar (ISR) to determine the meridional component of the neutral wind was originally suggested by Vasseur [1969] and has been applied widely to the study of thermospheric dynamics at various radar locations, namely, Millstone Hill, Sondrestrom, Arecibo, EISCAT (Scandinavia) and Saint Santin. The technique involves determination of the ion velocity parallel to the magnetic field at a particular height from field aligned radar Doppler measurements. The drift velocity of the ions is due to the combined action of neutral winds, plasma diffusion and electric fields. With the measurements made simultaneously on F region plasma parameters, subtraction of the diffusion velocity component from the observed drift velocity and projection of the residual velocity in the horizontal plane yield a measure of the meridional wind.

The major source of uncertainty in this method comes from limited knowledge on the ion-neutral collision frequency which determines the diffusion coefficient. The
statistical errors due to the measurement of the various parameters lead to an error estimate of 20–30 m/s in the neutral wind at $F$ region heights [Oliver and Salah, 1988]. During disturbed periods, the uncertainty may be as high as 100 m/s. Such large uncertainties lie in the neutral composition being used to compute diffusion velocity.

Contributions from the ISR at Arecibo, Puerto Rico, to the thermospheric dynamics are many. Nelson and Cogger [1971] first made use of the data on ion velocity and observed a large drop in the height of the $F$ layer, at Arecibo, after midnight. With the technique of determining meridional winds from ISR data being well established, Behnke and Harper [1973] began the study of ion-neutral interactions and examined the morphological features of the dynamic $F$ layer over Arecibo. Harper [1973] in his pioneering work, inferred large equatorward meridional wind speeds in the 2100–2400 h sector, which decreased rapidly after midnight and reversed sometimes near 0200–0300 h sector. This led Behnke and Kohl [1974] to conclude that the midnight collapse of $F$ region observed at Arecibo, is caused by the post-midnight poleward winds dragging the $F$ layer ions down the magnetic field lines. This picture evolved slowly with measurements from other techniques rapidly coming up, and has proved to be important in maintaining or depleting the nighttime ionosphere, particularly at magnetic midlatitudes, by the meridional winds. To a lesser extent, these winds influence the $F$ layer densities at low latitudes as well.

(ii) Inferences from satellite measurements of airglow emissions

Utilizing the greater spatial coverage by a satellite, Bittencourt et al. [1976] used the OGO 4 tropical nightglow data on 1356 Å and 6300 Å atomic oxygen airglow to infer meridional wind velocities. They have made use of the fact that the morphology of the tropical emissions is associated with processes such as Appleton
anomaly which cause the $F$ layer to be situated at different heights and with different values of electron density in different locations. From the latitudinal asymmetry present in the tropical emissions, they have deduced meridional wind patterns over low latitudes. The difference in the height of the $F$ layer at northern and southern conjugate points is directly related to the sum of the components of wind velocity in the magnetic meridian at the conjugate points. In a subsequent study, Bittencourt and Tinsley [1977] have analyzed simultaneous measurements of the electron density, O$_2^+$ density and zenith 6300 Å column emission rate obtained from the AE-C satellite, to infer meridional wind speeds. Their results agreed with the models of the global thermospheric wind system. The number of inferences made in this method is limited because of the requirement that there exist asymmetries about the magnetic equator in the ionospheric plasma parameters.

The other work reported in the literature concerns the AE-E measurements of 6300 Å emission by Barrage et al. [1990]. The satellite brightness measurements of 6300 Å emission are first subjected to a two-dimensional inversion technique to obtain altitude profiles of the volume emission rate. The volume emission rates are then converted to electron density profiles, which provide estimates of the $F_2$ layer height. In conjunction with MSIS-86 model to obtain neutral atmospheric densities, the layer peak heights are used to derive meridional winds. They have taken into account the effect of zonal electric field by incorporating typical values obtained by the ISR at the equatorial station, Jicamarca. They have thus demonstrated the utility of the AE-E 6300 Å data base for the investigation of the tropical neutral wind pattern as a function of latitude, longitude, local time and season. The main results arrived at by them are as follows.

In the Indian sector, the derived winds are almost exclusively northward over the entire latitude and local time range, indicating a clear summer to winter flow. A
distinct abatement of northward wind in the midnight sector is observed during this summer season in the southern hemisphere. The results obtained by them compare reasonably well with both HWM-87 and TGCM models, which however, do not predict the midnight abatement feature observed both in Indian and Pacific sectors. The main limitation in this method is the inability to derive meridional winds very close to the magnetic equator. Since the layer displacements due to meridional winds are very small near the equator, large uncertainties arise in the derived winds.

(iii) Meridional winds derived from ground-based ionosondes

This is the third method which is widely used by various workers and is based on the servo model concept of Rishbeth [1967]. Since the work of Ganguly et al. [1980] who showed that the servo model method can be successfully exploited to understand the ionosphere over Arecibo, and demonstrated its behaviour to be in accordance with the model prediction, various workers have attempted to determine the meridional wind field from the vast amount of ionospheric data regularly archived in the existing ground-based ionosondes [Buonsanto, 1986; Miller et al., 1986; Forbes et al., 1988; Krishnamurthy et al., 1990]. These workers have proved on and often that besides complementing the capabilities and shortcomings of other methods of determining neutral atmospheric parameters, ionosondes can provide data to validate and cross-check with their own, the outputs available from various numerical models.

The use of ionosonde data to infer meridional winds has attracted attention because of the recognition that the $F$ layer peak height $h_{\text{max}}$, which is determined by diffusion and chemistry, and is dependent on magnetic latitude (Chapter 4, section (4.3)), undergoes displacement ($\Delta h_{\text{max}}$), which is nearly linearly proportional to the meridional component ($U$) of the neutral wind [Rishbeth and Barron, 1960; Hanson and Patterson, 1964; Rishbeth et al., 1978]. If the balance height can be computed from a numerical model with realistic specifications of photochemistry and diffusion,
and with the assumption that the effects of electric fields are unimportant, then this method allows one to estimate meridional winds. This approach has been adopted by Miller et al. [1986], who found reasonable agreement of their estimates with other independent measurements available.

The method of Miller et al. [1986] involves the use of an ionospheric model to develop a relationship between the meridional wind and $h_{\text{max}}$. This is done by modeling the F layer at a few wind speeds. The meridional wind is then derived by a comparison of the modeled and measured layer heights. They have obtained better values of $\alpha$, the constant of proportionality, from the ionospheric model at various neutral winds by a linear regression through values of $h_{\text{max}}$. Fig. 5.2 shows the comparison of their model results for Arecibo with Fabry Perot measurements and those calculated from ion diffusion velocities obtained by the incoherent scatter radar at Arecibo. The three results show agreement to about ~ 30 m/s all through the night. However, as seen in the figure, the difference between their results and those derived from ISR data is about 50 m/s at midnight when the wind speed is rapidly decreasing.

Forbes et al. [1988] gave a parameterization of the linear dependence that exists between $\Delta h_{\text{max}}$ and $U_p$. This study was taken up by them as part of an attempt to delineate the effects of latitudinal penetration of ionospheric storms. Buonsanto [1986, 1988] made use of the servo equations of Rishbeth [1967] and Rishbeth et al. [1978]. Rather than computing $\alpha$ by modeling the F region at a few wind speeds as was done by Miller et al. [1986], he has used the MSIS-86 model to determine $\alpha$ through its dependence on the diffusion coefficient and the neutral scale height.

Krishnamurthy et al. [1990] made use of the ionosondes available in the equatorial region of Indian sector to derive meridional winds. The ionosondes are located at Trivandrum (8.6°N, 77°E, geographic; 0.8°S dip), an equatorial station, and at
Fig. 5.2. Comparison of the derived wind using ionosonde data with the direct measurements of meridional wind over Arecibo using Fabry Perot Spectrometer and the inferred wind ($U_B$) made from ion diffusion velocities obtained by the incoherent scatter radar at Arecibo. (after Miller et al., 1986)

Fig. 5.3. Nocturnal variation of meridional wind obtained by Krishnamurthy et al. [1990] for the low latitude station, Sriharikota (SHAR) in the Indian sector. Ionosonde data from Trivandrum, a station situated at the magnetic equator and from SHAR ($I = 10^\circ$N), slightly away from the equator, have been made use of for this purpose.
Sriharikota (SHAR) (13.7°N, 80.2°E, geographic; 10°N dip), a station slightly away from the dip equator. Assuming that $\mathbf{E} \times \mathbf{B}$ drift is the same at both stations, the vertical drift at SHAR is given by

$$v = v_D \cos I - U \cos I \sin I - w_D \sin^2 I$$

where $U$ is the meridional wind, $w_D$ is the plasma drift due to diffusion, $v_D$ is the vertical drift over the magnetic equator and $I$ is the dip angle. This is essentially subtraction of drifts due to electric field and diffusion from the observed drifts at a place where the meridional wind is to be estimated, to yield the drift due to wind. The rate of change of $h'F$, the virtual height of the $F$ layer, at both stations is assumed to represent the drift velocity of the plasma. They have incorporated the effect of apparent drift due to chemical loss. Fig. 5.3 shows the average nocturnal variation of $U$ (positive poleward) in the September month of 1988 as reported by Krishnamurthy et al. [1990]. Recently, Sekar and Sridharan [1992] validated the above method and showed that the existing ground-based ionograms can be effectively used for deriving meridional winds during nighttime at least in the Indian sector. They have utilized the bottomside ionograms available from SHAR and thermospheric and ionospheric data obtained from a rocket experiment conducted in 1982 at SHAR, wherein high altitude vapour clouds were released and the winds were estimated directly by ground photography [Raghavarao et al.; 1987].

Buonsanto et al. [1989] have discussed in detail the assumptions and uncertainties involved in obtaining meridional neutral winds from ionosonde data. The greatest uncertainty in the servo model derived winds is the uncertainty associated with ion-diffusion coefficient. How close is the ion-diffusion coefficient (ion-neutral collision cross-section, as some authors prefer) determined by the currently available methods to the actual value is a reigning controversy. At night, when there
are few collisions between ions and neutrals, the diffusion coefficient is larger, and the uncertainties in the MSIS model parameters translate into greater uncertainty in the diffusion coefficient, and therefore in the estimated wind. This effect becomes pronounced during geomagnetic storms, when there arise large changes in neutral composition and temperature [Prolss, 1987]. Buonsanto et al. [1989] found it necessary to tune the servo model by changing the empirical constant ‘c’ appearing in the servo equations, to obtain a better agreement between the winds derived from the servo model and those derived from incoherent scatter velocity measurements. They demonstrated possible deviations the derived winds can undergo from incoherent scatter measurements when the allowed composition changes during a severe magnetic storm are incorporated in the servo model.

Another source of uncertainty lies in the electric field whose effects are not included by any of the research workers. Miller et al. [1987] showed that for quiet or moderate geomagnetic conditions, the effect of electric fields on the neutral wind determination is generally small and usually smaller than the statistical uncertainties in the calculation. In the present work too, it is shown (Chapter 4, section (4.6)) that the effect of electric field in the low latitude thermosphere-ionosphere system is relatively unimportant considering the immediate response of the $F$ layer to changes in neutral temperature and meridional winds. However, it was proved by Buonsanto et al. [1989] that during disturbed periods, the effects of electric fields must be included in calculations of neutral winds obtained by the servo model method.

Finally, the effects of neutral temperature changes on the $F$ layer height are not incorporated by any of the workers who derive meridional winds from ionosonde and incoherent scatter line-of-sight ion velocity measurements. It is shown in Chapter 3 (section (3.5)) that the currently available neutral atmospheric models do not represent the variation of neutral temperature properly on many nights, especially in
the low latitude regions covered under the present study. This effect on the derived meridional winds is examined in one of the sections to follow. It is demonstrated that the changes in neutral temperature, in the wake of the well known geophysical processes such as equatorial ionization anomaly, equatorial spread $F$ and thermospheric and $F$ region storms, which at times act as sources of energy inputs, must be included in the servo model calculations to derive meridional winds.

5.5 Method adopted in the present study to derive meridional winds

As described in the previous section, all the methods, namely, those of Miller et al. [1986], Buonsanto [1988] and Forbes et al. [1988], employ the linear relationship between the $F$ layer height displacement and the meridional wind which causes it, to determine the latter parameter.

In the absence of electric fields, the poleward meridional wind

$$U_p = \frac{h_o - h_{\text{max}}}{\alpha} = \frac{\Delta h_{\text{max}}}{\alpha} \quad (5.2)$$

where $\alpha = \frac{2H^2 \cos l}{(k+1) D_m \sin \gamma}$. This follows from the linearization of the servo equation (4.17) and the elimination of rate of change of layer height, $dh/dt$, being a valid assumption in the $F$ region.

The method adopted in the present work is similar to the one by Buonsanto [1988]. Use of MSIS-86 model to determine the balance height $h_o$, in the absence of $U_p$, and the constant of proportionality $\alpha$ at that height $h_o$, and the $h_{\text{max}}$ determined by the true height reduction of ionograms, yields $U_p$, the neutral wind in the magnetic meridian.
The spectroscopically measured neutral temperature is adopted as a representation of $T_{\infty}$, the exospheric temperature, in the MSIS-86 model, to deduce the neutral densities of O, O$_2$ and N$_2$, and obtain the relevant recombination and diffusion coefficients ($\beta_m$ and $D_m$) at each height. The iteration method same as employed to estimate the $F$ layer peak height in Chapter 4 (section (4.6)), is used to determine $h_o$ from the servo expression (4.16). At this height, as already stated in Chapter 4, the diffusion and the loss due to recombination are of equal importance. The $F$ layer peak height ($h_{\text{max}}$), at this instant, is found from the scaled ionograms obtained by the ground-based ionosonde. Equation (5.2) then yields the meridional wind $U_p$.

The uncertainty for the derived wind is given by

$$\frac{\delta U}{U} = \left[\left(\frac{\delta h_{\text{max}}}{h_o - h_{\text{max}}}\right)^2 + \left(\frac{\delta h_o}{h_o - h_{\text{max}}}\right)^2 + \left(\frac{\delta \alpha}{\alpha}\right)^2\right]^{1/2}$$

similar to that derived for $h_m$ in Chapter 4.

An $h_{\text{max}}$ lower (higher) than $h_o$ implies poleward (equatorward) wind since the result of the wind has been to move the plasma down (up) the magnetic field lines.

It is useful to analyze the extent of linearity that holds between $U_p$ and $\Delta h_{\text{max}}$ (in equation (5.2)) before any estimate of meridional wind is made. This can be done by solving the servo equation for the vertical drift of ionization, which in our case has been assumed to be due to the neutral wind in the magnetic meridian. The result is

$$- U_p \sin I \cos I = \frac{D_m \sin^2 I}{2H} \left[\exp \left(\frac{(h_{\text{max}} - h_o)}{H}\right) - \exp \left(\frac{k(h_{\text{max}} - h_o)}{H}\right)\right]$$

To illustrate the response of the $F_2$ peak to different wind speeds and to examine the nonlinearity in (5.4), the servo equation has been worked out for the solar geophysical conditions as on 10 April 1991, and the result is shown in Fig. 5.4. The FP measurement in the premidnight hours on this night, yielded a neutral temperature of 1071 K, which determines the balance height in the absence of wind to be at about...
Fig. 5.4. The expected response of the $F$ layer peak height, $h_{\text{max}}$, to different wind speeds at the low latitude station, Mt. Abu. The servo equation (5.4) has been solved for different $F$ layer peak heights and the result is the non-linear curve. The dashed line represents the linear relation between the layer displacement and the meridional wind.
325 km. $D_m$ and $H$ in (5.4) are the parameters at this height $h_0$. Equation (5.4) is then solved for different $F_2$ peak heights to evaluate the corresponding meridional winds which cause the displacement. Also shown in the figure, is the result of calculations based on equation (5.2), i.e., the linear relation between $\Delta h_{\text{max}}$ and $U_p$. It is evident from Fig. 5.4 that the response of the $F$ layer is nearly linear (the deviation is less than 25 km) for equatorward wind speeds less than 200 m/s and for poleward wind speeds less than 125 m/s.

The assumption of linearity would then lead to an underestimate (overestimate) of equatorward (poleward) wind, which becomes significant if the layer gets displaced upward (downward) by more than 100 km (50 km). These numbers differ as the solar activity changes. For low solar activity period, it has been found that the slope of the linear curve decreases which implies the effect of nonlinearity on the derived winds is slightly reduced.

5.6 Results on derived meridional winds from low latitudes

The Fabry Perot Spectrometer which was commissioned at Mt. Abu (24.6°N, 72.7°E, geographic; 33°N dip) in 1985, has been yielding neutral temperatures since then and meridional wind measurements since the winter months of 1989. Since the required temperature stability of the etalon of about 0.1° C (which corresponds to an uncertainty in wind of about 10 m/s), was achieved only during December 1989, direct measurements of neutral winds are not available for earlier periods. Making use of the ionosonde data that are continuously being obtained from Ahmedabad (~ 2 degrees in latitude south of Mt. Abu) and the temperature measurements from Mt. Abu, meridional winds are derived for the period 1986-1989, using the method described in the previous section. Data belonging to quiet days only are taken up for the present.
study.

Seven examples are depicted in Figs. 5.5a to 5.5g, to illustrate the results. Two cases were considered for all the examples, one using MSIS-86 model temperatures and the other using measured temperatures, and the differences between them have been investigated.

Figs. 5.5a to 5.5d show the winds derived for four nights of data during October 1986. The agreement between the winds derived using measured $T_n$ and those derived with model $T_n$ is overall good except at times when the measured $T_n$ deviated considerably from the model $T_n$. These differences have been shown in Chapter 3 (section (3.5)). From equation (5.2), for a given $h_{max}$, the derived wind depends on $h_0$ and $\alpha$. Since the change in $h_0$ is more than that in $\alpha$ with height, the meridional wind increases in magnitude (more poleward or less equatorward) if the balance height $h_0$ determined by the measured $T_n$ is greater than that determined by model $T_n$. A striking example has been the result on 27 October (Fig. 5.5d). The deviation between the curves of about 30-50 m/s is seen throughout the period for which $T_n$ measurements are available. This implies an uniformly higher neutral temperature than what MSIS-86 model predicts for this night. Five out of seven $T_n$ measurements on this night were atleast 125 K more than the model values (as can be seen in Chapter 4, section (4.5)).

On 23 and 24, October (Fig. 5.5a and 5.5b), apart from differences in the range 50-70 m/s at certain times, the trend appears to be the same in both the cases for these days. On 26 October (Fig. 5.5c), the direction of the wind was opposite to each other for the two cases in the time period 2000-2130 h. The deviation was as large as 150 m/s. There happened to be a cross-over on this day implying that the temperatures in the time period 2000-2130 h had been less than the model values while those after about 2130 h had been more than what were predicted by the MSIS.
Figs. 5.5a,b. Comparison of derived meridional winds using spectroscopically measured temperatures and those derived using MSIS-86 model temperatures, for 23 (top panel) and 24 (bottom panel), October 1986. The method of deriving them is based on the servo model.
Figs. 5.5c,d. Same as Figs. 5.5a,b but for 26 October (top panel) and 27 October (bottom panel), 1986.
Figs. 5.5e,f. Same as Figs. 5.5a,b but for 23 November (top panel) and 1 December (bottom panel), 1986.
Mt. Abu/Ahmedabad

Feb. 13, 1988
Ap = 14

Wind using measured Tn
Wind using model Tn

Poleward wind (m/s)

Time (IST) Hrs.

Figs. 5.5g. Same as Figs. 5.5a, b but for 13 February 1988.
model (Chapter 3, section (3.5)). Despite small differences, for days 23 November and 1 December, 1986, the agreement was good (Fig. 5.5e and Fig. 5.5f). The poleward wind on 1 December need not be as large as depicted in the figure, since the poleward wind is always overestimated depending on to what height \( h_{\max} \) the layer is pushed below the balance level \( h_0 \). However, the equatorward wind is only slightly overestimated as explained in the last section. On 13 February 1988 (Fig. 5.5g), the differences between the two cases considered were significant (~ 150 m/s) at least at two different times (the differences in temperatures between the model predictions and the measurements can be seen in Chapter 4, section (4.5)).

The number of measurements on nights selected for this case study has been limited extending only upto midnight. This being a solar minimum period, the airglow intensity had decayed rapidly after about 2300 h leading to the rejection of emission profiles for the retrieval of Doppler information. With modifications to be made shortly in the optics that would enhance the flux gathered by the detector of the central-aperture scanned Fabry Perot, it is hoped that we would be able to retrieve the Doppler parameters even beyond midnight, even during periods of weak emission as one encounters during solar minimum.

The important results we arrive at, in the present study, are the following:

- The balance height \( h_0 \) of the \( F_2 \) peak, in the absence of winds, is determined by neutral temperature.

- It has been shown that the measured temperatures need to be incorporated in all calculations associated with the derivation of meridional winds from ionosonde data.
5.7 Nocturnal and seasonal variabilities of meridional winds

In order to understand the nature of the forcings which establish the neutral wind patterns, it is necessary to study the variation of thermospheric winds over different periods ranging from a few hours on a given night to a solar cycle. The amount of available data in this regard for the low latitude and equatorial regions is very less. As reviewed in section (5.3), Biondi et al. [1990, 1991] and Burnside and Tepley [1989] have attempted to understand the behaviour of thermospheric winds over Arequipa, Peru and Arecibo, Puerto Rico, respectively over a long period. In their analysis, they have made use of the fact that the solar radiation acts as a principal driving force that sets up the wind patterns, and obtained overall agreement between the expected and the observed wind variations.

Though averaging individual measurements over a month or a season masks the day-to-day variations as well as features such as gravity waves occurring on a shorter time scale, say, a few hours on any particular night, it still yields fruitful results for variations on a longer time scale. The results we presented in Chapter 3 (section (3.5)) on neutral temperature measurements, suggested that only about 65% of variations observed in them can be explained by the solar cycle dependence of thermospheric temperature and for the remaining, other physical processes must be operative which at certain times, control the temperature variations. In this section and the following chapter, we shall discuss these processes at large. Essentially, they compete with the solar forcing to establish the temperature and wind patterns, and the measurements reported herein probably are the result of this intimate competition within the solar-terrestrial environment.

As already mentioned in the last section, meridional winds have been derived
for the period 1986–1989 when both measurements in neutral temperature and ionospheric peak height \( h_{\text{max}} \) are available, using the servo model concept described in Chapter 4. It has been shown earlier that the balance height of the \( F_2 \) layer peak is determined by the neutral atmospheric parameters, namely, temperature and densities and that the meridional winds act as a perturbation on the equilibrium state of the \( F \) region. In the last section, we have seen that winds derived from model temperatures differ significantly at certain times from those derived using measured temperatures. This has been the case when we adopted the case study approach. However, for an average picture over a long period, say, a month, it is suffice that we make use of the model temperatures to derive meridional winds.

We present an example in Fig. 5.6 wherein the average winds obtained from the two different methods for the month of October 1986 are depicted. The spectroscopically measured neutral temperature and the ionospheric peak height \( h_{\text{max}} \) determined by the true height reduction of ionograms obtained from the ground-based ionosonde located at Ahmedabad, are made use of, at each instant, to estimate meridional winds, using the servo principle. MSIS-86 model exospheric temperature is also used to derive atmospheric densities and hence winds. The estimated wind values are sorted out in bins of half an hour duration and are then averaged for all the nights. The vertical bars (given only for winds derived using measured temperature) associated with each point represent the 1\( \sigma \) deviation among the individual estimated values. It can be seen that the winds derived using the atmospheric parameters determined by the MSIS-86 model for the two cases, namely, the measured temperature and the model temperature are within 1\( \sigma \) except for the large deviations near midnight. It is to be noted that the results presented in Chapter 3 indicated a large deviation in average temperature from the model prediction for this month (October 1986) after 2200 h. Since both equatorward winds and increase in neutral temperature act together in raising the height of the \( F_2 \) peak, the role of equatorward winds is magnified if one
Fig. 5.6. Average derived meridional winds over Ahmedabad for October 1986, using the ionospheric data. Wind derived using measured $T_n$ and those derived using model $T_n$ are compared.
uses the model temperatures which slowly decrease throughout the night. This effect tends to get nullified if we consider the average picture since the increase in measured $T_n$ is not observed on all nights and does not occur at the same time. Thus for gross, average, features, the winds derived using the atmospheric densities determined by the model temperature can be taken to represent the average nightly variations in thermospheric winds. However, for the individual case study approach, we stress the importance of observed neutral temperature and its variations which need to be incorporated in deriving meridional winds.

For the present study, we have also used the horizontal wind model (HWM) of Hedin et al. [1988] and the vector spherical harmonic model of Killeen et al. [1987]. The HWM is an empirical model based on wind data obtained from the AE-E and DE 2 satellites. A limited set of vector spherical harmonics has been used to describe the zonal and the meridional components of horizontal wind. A modified version of the model is now available [Hedin et al., 1991]. The VSH model is based on the output from a set of runs of the NCAR Thermospheric-Ionospheric General Circulation Model (TIGCM) [Dickinson et al., 1981] generated on the Cray supercomputer in Boulder, Colorado. An advantage of the VSH model is that the geophysical fields of interest can be obtained over a wide range of solar-geophysical conditions, even on a computer with modest storage requirements. These models are more commonly used nowadays among the upper atmospheric researchers, since they not only provide an understanding of the spatial and temporal variation of the geophysical parameters but also help in testing them against the direct measurements available and thereby aid further understanding.

The average derived winds and their variations for the seasons (post-equinox and winter) under study are depicted in Figs. 5.7a to 5.7d, the post-equinoctial months being October and November (belonging to years 1986 and 1988 respectively) and
Fig. 5.7a. Nocturnal variation of derived meridional winds over Mt. Abu/Ahmedabad averaged for October 1986. The horizontal wind model (HWM) prediction is shown for comparison.

Fig. 5.7b. Same as above but for November 1988.
Fig. 5.7c. Same as Fig. 5.7a but for January 1988.

Fig. 5.7d. Same as Fig. 5.7a but for February 1988.
the winter months being January and February (1988). Due to the limitations in the availability of ionospheric data, we have selected only four months for this exercise, the average number of nights for each month being only 6. The data belong to quiet geomagnetic conditions. For each of these months, the horizontal wind model is run, and the result is also depicted along with the average variation of estimated winds. The vertical bars represent the night-to-night variations in the derived winds.

Figs. 5.7a and 5.7b show the variation of average derived meridional winds for the months of October (1986) and November (1988). In the post-equinoctial period, the winds exhibit considerable deviations from the model predictions in October, 1986, while the agreement is reasonably good during November, 1988. During October, as discussed earlier, the equatorward wind becomes very large and shows maximum amplitude (~ 200 m/s) at about 2315 h. The wind shows a constant speed of ~ 50 m/s near midnight in the month of November.

Similar variations are seen in the winter months of 1988. The HWM yields small equatorward wind speeds of less than 30 m/s throughout the night, in January. A strong transequatorial flow (from the southern hemisphere) is expected to get superposed over the normal equatorward wind and the net result is a weak equatorward wind. Though a transition is seen from the poleward to the equatorward flow, the estimated wind magnitudes are significantly higher than the model values, near and after midnight. The smallest vertical bars indicate the sparsity of ionograms available during the particular time slot, while the largest error bars indicate the large day-to-day variations in the derived wind at 1σ level. The average picture that emerges for these two seasons is that there exists a poleward or a small equatorward wind in the early part of the night which then becomes an equatorward wind of large magnitude (150 to 200 m/s) near midnight. There are not much differences in the wind variabilities between the two seasons, the overall features remaining similar.
Before we discuss about the seasonal variations exhibited by direct measurements of meridional winds, a few comments assume significance regarding the results presented above. They are only a few examples selected during the solar minimum years 1986–1988 to show the sort of variations the meridional winds undergo for the two seasons (post-equinox and winter). Though it had been demonstrated earlier that during this solar epoch, the use of model temperatures instead of the measured ones does reasonably represent the actual conditions, these would not be valid during the solar maximum period when the swings in the observed temperature variations are significantly larger as has been shown in Chapter 3 (section (3.5)). This is illustrated by examples shown in Figs. 5.8 and 5.9. In Fig. 5.8, the direct measurements for 17 February 1991 are compared with winds estimated for the two cases discussed earlier. The estimates using measured $T_n$ are seen to follow closely the measured meridional winds than those using model $T_n$ implying the importance of temperature variations in influencing the derived winds. There has been a systematic difference in wind magnitudes between the two estimates in the time sector 2000–2200 h. While the estimated wind with measured $T_n$ determining the balance height $h_o$ does show the transition to poleward as exhibited by the direct measurements, the wind with model $T_n$ continues to blow equatorward. The large differences between both the estimates and the measurements in the late night hours are due to electric fields not being taken into account while deriving the meridional component of the neutral wind.

Fig. 5.9 depicts the average variations exhibited by direct measurements and the estimates based only on model temperatures for the month of December 1989. The vertical bars attached to each data point in the case of measured mean winds indicate the night-to-night deviations in the observations. On comparison between the two, significant differences are evident especially in the 2200–0000 h sector. The average observed wind had become poleward at about 2230 h whereas the derived wind remained equatorward with a magnitude of ~ 100 m/s. There are differences
Fig. 5.8. Comparison of meridional winds derived using ionospheric data for the two cases, namely, those with measured $T_n$ and those with model $T_n$, with spectroscopic measurements of meridional winds for 17 February 1991. The significant effect of measured neutral temperatures in influencing the derived winds may be noted.

Meridional winds over Mt.Abu (Dec.89)

Fig. 5.9. Comparison of derived winds based only on model temperatures with direct measurements for December 1989.
in the early morning hours as well. These examples strengthen the conclusion arrived at in the last section that the effects of temperature variations need to be incorporated while deriving meridional winds from the ionosonde data at least during periods of high solar activity.

Long-term variability of measured meridional winds

As had been mentioned earlier, the direct measurements of meridional winds are being made since the winter months of 1989. The required stability of the Fabry Perot etalon is about 0.1° C for useful data on line-of-sight winds. Though the thermal stability of the etalon has been achieved upto this desired limit, during certain campaigns, the drifts observed have been very large and the data belonging to these measurements are not considered for analysis. Even on nights when we could maintain the required stability, the intensity of 6300 Å happened to be too low during the early night period and exhibited fast variations during subsequent times calling for extreme care in selecting the line profiles for analysis. Due to these various reasons, we have a limited data base on direct measurements of neutral winds. In spite of this limitation, important results and conclusions have still been arrived at, and these will be discussed shortly.

Fig. 5.10 shows an example in which the measurements of meridional wind made on 28 December, 1989, from Mt. Abu are compared with the VSH and the HWM model runs. The VSH model prediction for this night indicates a poleward wind all through the night implying a transequatorial flow of air originating in the southern (summer) hemisphere. The HWM model predicts a small equatorward velocity (~30 m/s) around midnight. The measurements show a wavy pattern of a periodicity of about 4 hours, the magnitude and the direction differing significantly from the models. The possibility of sources of such large deviations of the observed values from the model predictions will be discussed later.
Fig. 5.10. Comparison of measured meridional wind with HWM and VSH model results for 28 December 1989.

Averaged meridional winds over Mt. Abu (Dec.89)

Fig. 5.11a. Nocturnal variation of measured meridional winds from Mt. Abu for December 1989. Individual spectroscopic measurements are plotted here. The continuous curve depicts the average values of data points put in slots of half an hour duration.
Averaged meridional winds over Mt.Abu (Feb.'91)

Fig. 5.11b. Same as Fig. 5.11a but for February 1991.

Averaged meridional winds over Mt.Abu (April '91)

Fig. 5.11c. Same as Fig. 5.11a but for April 1991.
We shall be deviating from the case study approach we have adopted till now, and look at the gross features indicating the response of the thermosphere to the underlying forcings. For such a study, it is necessary that the number of measurements is large enough to draw any meaningful statistical inference. Adopting a criterion of at least 30 individual measurements needed from a campaign, we have selected three months, namely, December (1989) and February (1991), and the equinox month of April (1991). The total individual measurements (data points) run up to about 140, the largest of 65 being in the December month. The method of averaging is the same as described earlier. The results are shown in Figs. 5.11a, 5.11b and 5.11c for these three months. Individual measurements are also shown so that the day-to-day variabilities become evident. A first look at these figures indicates that the measured winds are largely in the equatorward direction. Though the day-to-day variations are significant, still a systematic pattern emerges for each of these months.

The results for December 1989, are depicted in Fig. 5.11a. The average $A_p$ value during this moderately disturbed period had been 19, the lowest being 10 on 28 December and the highest being 26 on 22 December. There had occurred a geomagnetic storm on 29 December with $A_p$ value of 50. We have not considered the data belonging to this day, which incidentally happened to be the last day of the campaign during this month. For the remaining period, we can safely assume the geomagnetic activity to be moderate to low.

With the exception of large variability following sunset hours, the winds fit into a pattern during December 1989. From a large equatorward wind (of about 100 m/s), the average wind either approaches small equatorward or turns poleward between 2100 to 2300. From 2300 h onwards, the equatorward wind gains in magnitude and reaches its highest amplitude of $\sim 150$ m/s at about 0130 h. There is a tendency to become poleward at 0300 h after which the wind becomes equatorward once again.
Let us look into the wind pattern for the month of February (1991), depicted in Fig. 5.11b. This has been a geomagnetically quiet period (average $A_p$ value for the period of observations during this month being 10). The day-to-day variability is large for this month when compared to other months. However, a closer look suggests that it is not much different from the average picture that has emerged for the month of December. The mean wind in February is poleward around 2100 h which then gradually becomes equatorward. Since the largest changes in intensity occur around midnight and immediately after, the number of measurements is less during this period. However, there is a clear transition to poleward at about 0300 h, the trend being similar to the one that occurred during December. These might be the characteristics of a winter season.

The variation for the month of April (1991) is depicted in Fig. 5.11c. This also has been a geomagnetically quiet period (average $A_p$ value for the period of observations during this month being 9). In this month also, the winds tend to follow an established pattern. A large equatorward wind often occurs in the early night hours which then reduces in magnitude, sometimes becoming poleward at about 2300 h. At about 0300 h, the average wind reaches its highest amplitude of about 175 m/s after which the wind decelerates. An important point to be noted here is that the time at which the equatorward wind attains the maximum speed is ~ 2 hours later when compared to December and February months.

Even with a limited data base representing winter and equinoctial periods, there emerges systematic behaviour of thermospheric winds in these two seasons. In both the seasons, the wind makes its transition to poleward for a brief period at about 2300 h. In the winter months, the equatorward wind reaches its highest value at about 0100 h while during the equinox period, the maximum equatorward wind occurs at about 0300 h exhibiting a shift in phase of about 2 hours.
It is worth comparing the seasonal behaviour of winds over Mt. Abu with that at other latitudes. At Cachoeira Paulista (22.7°S, 45°W, geographic; 16°S dip latitude), Brazil, during winter months, the wind velocities are poleward early in the night, reduce to zero around midnight and change to poleward again. During the spring-equinox period, the average winds are largely equatorward in the pre-midnight hours which then reduce in magnitude to zero around 0100–0200 h [Sahai et al., 1992]. They report that the wind patterns are similar to those predicted by HWM-87 and HWM-90 and apart from some discrepancies in absolute magnitudes, the trends appear to agree. Considering that their observing location is situated within the purview of equatorial ionospheric and South Atlantic magnetic anomalies, their result that the wind patterns are in overall agreement with HWM models needs to be considered carefully. Biondi et al. [1991] have compared the winds over Arequipa (16.5°S, 71.5°W, geographic; 3.2°S dip latitude) with those observed at Arecibo (18.6°N, 66.8°W, geographic; 30°N dip latitude). The early night winds during winter at both the locations are weakly poleward, reversing direction in the later part of the night, conforming with the expected nocturnal pattern. At Arequipa, the data during equinox periods indicate a large equatorward wind in the early part of the night which then dies down to zero around midnight and again becomes weakly equatorward in the late night. Biondi et al. [1991] have compared the winds over Arequipa with the airglow data of Burrage et al. [1990] and the findings are in agreement with the sense of a general transequatorial flow at the solstices. From these results it is difficult to construct a coherent picture for all the low latitude stations located at different longitudes.

Biondi and Sipler [1985] observed some unusual features in the flow of thermospheric winds and the temperatures over Natal (5.9°S, 35.2°W, geographic; 6.4°S dip latitude) identical to those observed earlier in midlatitude studies [Hernandez, 1982c]. On a particular night of 26 August 1982, they have observed convergence in
The meridional wind between 2100 to 2330 UT accompanied by a downward wind and an associated temperature increase. The geomagnetic activity had remained low and hence their effects can be ruled out. Owing to the renewed interest by the work of Raghavarao et al. [1991, 1993], the first direct experimental evidence for the presence of Equatorial Temperature and Wind Anomaly (ETWA) colocated with the EIA, the EIA and its associated effects on thermospheric circulation have become relevant. The meridional wind measurements from Mt. Abu appear to provide credence to the hypothesized circulation cells associated with ETWA getting set up. We shall be discussing this process and the relevance of our measurements in detail in the next chapter.

To conclude,

- The predictions made by the wind models, namely, HWM and VSH, do not agree with the observations from low latitudes.

- The observed features are considerably different from those reported from other latitudes showing significant longitudinal difference.