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LECTURE NOTES ON MESOSPHERIC WINDS AND TURBULENCE

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Lecture notes on Mesospheric winds and turbulence

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Introduction:

Recent efforts of many scientists, making use of the potential of ground based radar measurements in examining the low frequency motions and evaluating their contributions to the observed dynamical behavior of the mesosphere and lower thermosphere, offered a better understanding of the mesosphere [*Vincent and Lesicar*, 1991; *Harris and Vincent*, 1993; *Vincent*, 1993; *Fritts and Isler*, 1994; *Palo and Avery*, 1995]. However, there are many unsolved problems that need to be quantitatively understood, to name a few, effects of long period waves (equatorial mesosphere) on the background mean flow, gravity wave –tide-planetary wave interactions and their effects and contribution of tides and gravity waves to the momentum budget of the mesosphere. This calls for a number of radar measurements that are separated in latitude and longitude in this region. Before going into details of the parameters which radar measures, we will see a brief history of its development.

A brief History of development of radars:

History of radar studies of the clear atmosphere has roots in 1930s and has been well reviewed by *Hardy and Gage* [1990]. In the mid 1930s, the 10-200 m wavelength radars, used for ionospheric studies, detected reflecting layers in the troposphere where it was known that ionized regions could not exist. Theories were developed for the scattering of electromagnetic radiation in the troposphere by *Pekeris* [1947], *Booker and Gordon* [1950], *Tatarskii* [1961] and many others. In the 1960s, a new generation of high-power, high-frequency radars were developed for investigating atmospheric structures. In 1970s another type of radars using VHF band frequencies and large phased array antennas were developed, primarily for ionospheric research. However, *Woodman and Guillen* [1974] in their pioneering work at Jicamarca, Peru observed echoes from the lower atmosphere and mesosphere. This work led to the concept of Mesosphere-Stratosphere-Troposphere (MST) radar, and this class of radars has come to dominate the atmospheric radar scene over the past decade. The MST (Mesosphere-Stratosphere-Troposphere) technique, which uses sensitive radars in the frequency range 30-3000 MHz to examine the optically clear atmosphere below roughly 100 km, can be considered as having evolved from the pioneering work of *Woodman and Guillen* [1974]. The origin of the echoes in these regions was determined to be Bragg scattering from refractive index fluctuations associated with turbulent irregularities of scale sizes nearly half of the radar wavelength. Because such turbulence is virtually ubiquitous in the atmosphere and it can be used as a tracer of the clear air motion, provided it moves with the mean flow.

Studies of *Woodman and Guillen* [1974] triggered the evolution of a new generation of VHF/UHF radars for atmospheric research. The first one of the new radars was built near Boulder, Colorado at Sunset [*Green et al.*, 1975] and the SOUSY (SOUnding SYstem) VHF radar in West Germany followed soon after [*Czechowsky et al.*, 1976]. These radars were successfully operated and the wind measurements by Sunset radar are well correlated with that observed by the rawinsonde [*Van Zandt et al.*, 1975]. Layered structures in the radar echoes were detected with SOUSY-VHF radar [*Rottger and Czechowsky*, 1977] and with the Sunset radar [*Gage and Green*, 1978; 1979]. Since then a number of radars capable of sensing turbulence in the neutral atmosphere have been built and operated throughout the world. More background information can be found in the publications referenced below in which technical details and the first experimental results from some of the well known radars were first described. The Poker Flat radar in Alaska has been described by *Balsley et al.* [1980, 1981]; the MU radar near Kyoto, Japan by *Fukao et al.* [1985a, b]; the Buckland Park radar near Adelaide, Australia by *Vincent et al.* [1982, 1987]; the Arecibo 50 MHz radars by *Rottger et al.* [1981, 1986]; and the Chung-Li radar in Taiwan by *Brosnahan et al.* [1983]; *Chao et al.* [1986], Equatorial radar at Indonesia by *Fukao et al.* [1990] and the Indian MST radar at Gadanki, India by *Rao et al.* [1995]. *Hocking* [1997] described some of the existing MST/ST radars in his review article.

The basic parameters measured by MST radars are echo strength, Doppler shift and Doppler width. Atmospheric variables deduced from continuous measurements of these parameters include mean winds, planetary waves, tides, gravity waves, turbulence structure and atmospheric stability [*Green et al.*, 1979]. The unique feature of MST Radar is its capability to measure the vertical wind. *Miller et al.*, [1978] were the first to report mesospheric vertical velocity measurements by the VHF radar at tropical latitudes.

High time resolution vertical wind measurements contain a great deal of information on gravity wave activity, while the long terms mean vertical velocities are potentially used to study the large-scale circulation pattern. Vertical winds have been inferred from horizontal wind convergence and continuity arguments. Knowledge of vertical wind field is of considerable importance in understanding exchange processes between different spheres, large-scale circulation patterns (e.g., Hadley and Walker circulations) and wave transport phenomena.

Wind measurements

Before going into details of wind measurements by radar techniques, it is desirable to see how these measurements are mad. Two methods are generally used for measuring velocities with UHF/VHF radars. These are the Doppler Beam Swinging (DBS) and the Spaced Antenna Drift (SAD) methods. The two methods are shown schematically in Figure 1.

The DBS method uses two or more narrow beams pointed in different directions, one at zenith to measure the vertical motion directly and other beams at some small zenith angle. If the contribution of vertical motion to the off-vertical radial velocities is negligible, the wind velocity vector can be determined by measuring the Doppler frequency of Bragg scatter echoes. There are several reasons for choosing the zenith angle of the antenna beam to be as small as possible. The wind field is more spatially uniform for smaller separations of the probed radar volumes. Further, decreasing the zenith angle increases the effective antenna area, increasing the gain and the sensitivity of the system. Another advantage in using small zenith angles is that the echoes may be several orders of magnitude greater than for off-vertical angles of 10^0 or more. However, care has to be taken, because the velocity estimates can be biased due to the apparent-look-angle effect if aspect sensitivity is important [*Rottger*, 1980].

The SAD method uses a vertical-beam antenna for transmission and three or more horizontally spaced, vertical beam receiving antennas. The scattered radiation forms a diffraction pattern on the ground, and the motion of the pattern on the surface is found from the cross-correlation of received signals for the receiving antennas [*Briggs*, 1968, 1977; *Rottger and Vincent*, 1978]. Using simple geometry, it may be demonstrated that the drift velocity of the diffraction pattern across the receiving array is twice the horizontal wind velocity in the radar scattering volume.

The advantages and disadvantages of these systems are discussed in detail by *Rottger* [1981]. Aspect sensitivity argues in favour of the spaced antenna method since it uses vertical beam antennas and, as a result, an increase in the signal-to-noise ratio is usually gained. Further, antenna side lobe effects are less important and the fixed vertical beam positions are technically much easier to handle than steerable antenna beams. On the other hand, DBS methods can often achieve greater spatial and temporal resolution, especially for single-beam studies [*Hocking*, 1997]. A detailed intercomparison of wind velocities measured with spaced antenna and the Doppler method has been carried out by *Vincent et al.* [1987], who also made comparisons with 80 radiosonde profiles and found a good correlation between these measurements.

Wind measurements over tropical latitude:

The equatorial atmosphere within a few degrees of the equator is virtually devoid of Coriolis forcing. As a result, a separate class of relatively long period waves (Kelvin waves and mixed Rossby gravity waves) exist in this region. The dynamics of the equatorial atmosphere is thought to be strongly controlled by the evolution and breakdown of these waves. Waves of this kind are believed to contribute significantly through the complex wave mean flow interactions to some of the observed variations in the dynamical parameters, for example, the Quasi Biennial oscillations (QBO) [*Veryard and Ebdon*, 1961; *Reed et. al.*, 1965] and the Semi Annual Oscillations (SAO) [*Hirota*, 1978; *Dunkerton*, 1979], in the middle atmospheric mean zonal wind. For the study of these waves, satellite measurements with the Limb Infrared Monitor of the Stratosphere (LIMS) experiment on board on Nimbus-7, the Microwave Limb Sounder (MLS) and High Resolution Doppler Imager (HRDI) on board on Upper Atmospheric Research Satellite (UARS) have added new dimensions to our current understanding of the long period equatorial wave propagation in the middle atmosphere [*Salby et al.*, 1984; *Deluisi* *and Dunkerton*, 1988; *Hitchmann and Leovy*, 1988; *Canziani et al.*, 1994; *Burrage et al.*, 1996]. Ground based radar measurements in the equatorial region have yielded useful information on the propagation characteristics of the long period waves [*Vincent*, 1993].

In early 1990s, observations with meteor radar and a new generation Partial Reflection Radar (PRR) were initiated at Christmas Island (2^0N, 157^0W) in the Central Pacific and preliminary results were reported by *Avery et al.*, [1990] and *Vincent and Lesicar*, [1991]. Later using long-term observations, *Eckermann et al.*, [1997] have studied intraseasonal wind variability in the equatorial mesosphere and lower thermosphere over Christmas Island in the central pacific. Another PR radar became operational at Hawaii (22^0N, 159^0W). In the Indian zone, a meteor radar operated from Trivandrum (8.5^0N, 77^0E) provided useful results on the low-latitude dynamical behavior of the atmosphere in the meteor region (80-105 km) [*Raghava Reddi and Ramkumar*, 1995; 1997].

Some of the tropical features in winds are discussed briefly here. Figure 2 shows the zonal component of winds over Tirunelveli (8.7^0N, 77.8^0E) [*Rajaram and Gurubaran*, 1998]. From the figure it is clear that westward flow in the equinoxes, eastward flow during Solstices and downward trend on many occasions. All these features reveal Semi Annual Oscillation (SAO) in the zonal wind. Other important feature is the inter-annual variability in the enhanced westward flow during spring equinox at mesopause region during 1993 and 1995, which is nothing but due to Quasi-biennial Oscillation (QBO). Figure 3 shows the meridional component of winds over the same altitude. It is clear from the figure that meridional component of wind is towards pole ward motion during winter and equator ward motion during summer.

The dominant motions in the equatorial mesosphere are depicted in the Figure 4. From the figure it is clear that semi-annual oscillation is the dominant mode observed in zonal wind. The first harmonic corresponds to the period of QBO. Below 82 km its amplitude is low (3 ms^{-1}) and grows upto 8 ms^{-1} at 86 km. The annual oscillation attains peak values at lower heights and tends to show an increase beyond 94 km. The third (period of 8 months), fifth (period of 5 months) and sixth (period of 4 months) harmonics reach peaks at 86 km with amplitudes in the range of 5-7 ms^{-1}. The higher harmonics are observed to be less significant. Breaking of the upward propagating gravity waves has been suggested to be the causative mechanism for the SAO [*Dunkerton*, 1982].

Meteor and Partial Reflection radars are widely used to measure the mesospheric and lower thermospheric motions [*Vincent*, 1993; *Raghava Reddi and Ramkumar*, 1997; *Rajaram and Gurubaran*, 1998; *Gurubaran and Rajaram*, 1999]. These MF and Meteor radars have capability to monitor the Mesosphere and Lower Thermosphere (MLT) winds on a continuous basis but from 80 km onwards with more reliable though it starts from 72 km with less reliability. Moreover, many models have been developed based upon rocket, satellite and various ground-based instruments. For example, CIRA 86 model contains monthly tabulations of zonal mean wind from 0 to 120 km [*Fleming et al.*, 1990] derived from a tropospheric climatology by *Oort* [1983] and use of the gradient wind approximation with the temperature and pressure tables derived from satellite remote sensing data by *Barnett and Corney* [1985] for the middle atmosphere and the mass spectrometer and incoherent scatter (MSIS-83) empirical model [*Hedin*, 1983] for the thermosphere. These model-derived winds are not suitable for all latitudes [*Manson et al*, 1991] especially for tropics [*Hedin et al*, 1996]. Moreover, many satellites were designed to study the winds. For example, HRDI on UARS is designed to measure horizontal winds in the mesosphere and lower thermosphere (50-115 km) and in the stratosphere (10-40 km). The main aim of this satellite is to improve the understanding of the global atmospheric changes and to develop a comprehensive global climatology of the winds in the mesosphere. Using this technique, extensive studies were made by *Khattatov et al.*, [1996]; *Geller et al.*, [1997]; *Hagan et al.*, [1997]; *Khattatov et al.*, [1997(a)]; *Khattatov et al.*, [1997(b)]; *Yudin et al.*, [1998]. These winds are extensively compared with other measurements [*Hasebe et al.*, 1997] over the globe at various latitudes by *Burrage et al.*, [1996] and found a very good to fair comparison at all latitudes except tropics. For example, Figure 5 shows a comparison of MF radar measured winds HRDI measured winds over various latitudes. From the figure it clear that at almost all the locations, radar derived winds and HRDI measured winds are matching well except with that of Christmas Island MF radar which is a tropical station. A large discrepancy can be seen between these two measurements. Similar observations are also noticed between Indian MST radar winds and HRDI measured winds, which is shown in Figure 6. Earlier MST radar derived winds are compared with the Partial Reflection radar (PRR) located with Tirunelveli (8.7^0N, 77.8^0E) (which is also a tropical station) which is 500 km away from MST radar site and found a good comparison between these two ground based instruments (see Figure 7 for more details).

Inorder to produce reasonable results from these models and satellites, it is necessary to incorporate dissipation and momentum deposition by upward travelling gravity waves (and to a lesser extent, planetary wave and tides). The high time resolution afforded by continuous observations using a near equatorial network of MST [Mesosphere, Stratosphere and Troposphere] radars should be extremely beneficial in study the evolution of these waves. These MST radars can be effectively used to overcome the difficulties of other techniques especially between 65-85 km height region inspite of some limitations [*Rottger et al.*, 1983]. Using the MU [Middle and Upper atmosphere] radar, *Nakamura et al.*, [1996] and *Namboothiri et al.*, [1999] have studied interannual variability of mesospheric mean winds over Shigaraki, Japan using large database (for about 10 years). In the tropical latitudes, VHF radar observations of wind velocities using SOUSY radar at Arecibo (18.35^0N, 66.75^0W) were reported by *Fukao et al.*, [1979]; *Rottger et al.*, [1983]; *Ruster et al.*, [1986], at Jicamarca (11.95^0S, 76.87^0W) by *Maekawa et al.*, [1986] and at Chung-Li by *Chu et al.*, [1995], using small data sets. Such a network of single point measurements would be highly complementary to the existing satellite database, which supplies gross features of the geostropic wind field and other parameters on a global scale. Before going into the details of the observed winds and its comparison with various techniques, it is better to discuss the systematic errors in estimating the winds using the radar technique at mesospheric heights.

Strengths and limitations of VHF radar measurements of mesospheric winds

The strengths and limitations of MST radar measurements of mesospheric winds have been discussed in detail by several authors [*Rottger et al.*, 1981; *Nakamura et al.*, 1996; *Hocking*, 1997]. Since the mesospheric echoes are greatly influenced by the presence of electron density fluctuations, which is a daylight phenomenon, radar observations are confined to day hours only. This involves difficulty in performing harmonic analysis to include diurnal and semidiurnal components. The observed winds by the simple averaging will be biased by the tidal variations. If this bias is smaller it is worth even if the daytime winds are averaged simply and can be utilized as mean winds. Another limitation is that the turbulence is not uniformly distributed over the entire mesosphere rather it occurs in the form of layers. Moreover, the echoes are highly intermittent in space and time. These echoes are confined to a few kilometers mostly in 70-80 km height region. Figure 8 shows the echo characteristics, which is taken from Indian MST radar observations on 23 March 1999. From the figure it is clear that more intense echoes will occur in the region between 70-80 km. Only for a few cases one could be able to see the echoes for the entire altitude region between 65-85 km. One more limitation is that the observations are limited on average for a few days in a month. Mean winds deduced by averaging for limited days do not always represent the seasonal variation of the mean wind. To overcome this problem, the data is to be averaged by collecting many years of data. Besides these limitations, VHF radar has several advantages over other techniques. MST radar measurements have sufficient signal-to-noise ratio in the height region of 65-85 km, which is not covered by the meteor radars. Moreover, the Partial Reflection Radars cannot reach these lowest heights. In this regard, it is worth having MST radar observations as useful complementary information on winds at these lower heights [*Rottger*, 1980; *Rottger et al.*, 1981].

By over coming these difficulties *Nakamura et al.*, [1996] and *Namboothiri et al.*, [1999] have successfully estimated mean winds over Shigaraki, Japan using MU radar. Following similar procedure mean winds are estimated using Indian MST radar over Gadanki and the results are presented in Figure 9. From the figure it is clear that all the features of tropical mesospheric winds as observed by *Rajaram and Gurubaran*, [1998] is noticed in this case also. The observed winds are also compared with the winds measured by MF radar over Tirunelveli and is already shown in Figure 7. The comparison is found to be good in the zonal component and some discrepancies in meridional component.

Similarly comparison with CIRA-86 model is also shown. In this case it is found that model-derived winds are showing large values when compared with the winds derived by ground-based instruments.

Wind measurements over mid latitudes:

Wind variability over mid and polar latitudes varies from that of tropical latitudes. Figure 10(a) and 10 (b) shows the mean wind variations of the zonal and meridional component respectively observed over Shigaraki, Japan (34.9^0N, 136.1^0E) using data collected by MU radar for a period of about 10 years [*Namboothiri et al.*, 1999]. Here we are just giving the example how mid latitude wind filed vary. From the figure it is clear that during summer, the flow is towards westward (easterly) and during winter flow is towards eastwards (westerly). A clear annual oscillation in zonal component is noticed which is entirely different from tropical latitudes in which a semi annual oscillation is seen. The values during the summer months (westward) reached upto 45 ms^{-1} at the height of 70km. The eastward flow attained a maximum value of about 40 ms^{-1} during winter. At mid-latitudes, the mean wind patterns during solstices are driven by pole to pole differential solar heating and the Coriolis force acting upon the resulting meridional circulation leads to easterlies during summer and westerlies during winter in the mesosphere [*Andrews et al.*, 1987]. The feature of QBO is also noticed by them even in mid altitudes which is a tropical phenomenon (see Figure 11). This figure shows the duration of summer westward flow, in approximate months from 1986 to 1995 at two heights, 70 and 75 km. The figure indicates a periodicity close to two years and this feature is almost repeated at the heights. This might be due to QBO, which is defined as direction of winds at 40-50 hPa over the equator is commonly taken to define the phase of QBO as westward or eastward. There is a little confusion regarding the penetration of equatorial QBO effects to mid -latitude mesospheric heights. There are also evidences [for example *Burrage et al.*, 1996] that the QBO can extend upto 30^0 in both the hemispheres.

The meridional component of wind shows a complicated structure. In general, northward winds are dominating at most of the height regions and months, except for a strong southward flow from May to the middle of August at above 70 km and a few other intermittent bands of southward motion in this height range.

Wind measurements over polar latitudes:

Figure 12 shows a typical example of zonal and meridional component of winds over Hally, Antartica (76^0N, 26^0W) [*Charles and Jones*, 1999] using Imaging Doppler Interferometer (IDI). From the figure it is clear that over polar latitudes there is westward flow during summer and eastward flow during winter below about 85 km. Above 85 km the directions are reversed with eastward flows in summer and a small westward flow during the middle of winter. The clear annual oscillation in zonal component can be noticed at mesospheric heights. The peak magnitude observed in the zonal component is about 10 ms^{-1}. The meridional component shows equatorward flow during summer between 80 and 95 km and poleward flow above and below. The amplitudes of the mean meridional wind are about 5-10 ms^{-1}. In winter, the meridional circulation shows a clear reversal with height with a small but clear poleward flow below 80 km and stronger equatorward flow above.

Mesospheric studies using combined radar and Lidar techniques:

The important aspect that MST radar technique can provide is that to study the effect of small-scale motions (gravity waves, turbulence) on the large-scale circulation of the atmosphere. The influence of small-scale motions on the general circulation is known to be appreciable, particularly in the upper mesosphere [*Rottger*, 1987; *Hauchecorne et al.*, 1987; *Fritts*, 1989; *Fritts and Lu*, 1993; *Eckermann and Marks*, 1997; *Liu et al.*, 1999]. Simple models that predict middle atmospheric structure and zonal mean flow are incapable of reproducing either the measured temperature profiles or the zonal mean flow. For example, all models assume a regular decrease of temperature from the stratopause to mesopause while the observations using Rayleigh Lidar (Light Detection and Ranging) measurements display quite frequently a secondary maximum of temperature between 70-75 km height region. Figure 13 shows the mean temperature profile derived from Nd: YAG lidar locater at Indian MST radar site. From the figure it is clear that a secondary maximum (assuming mesopause as first maximum) of temperature is clearly seen with 30 K enhancement in temperature from the mean. Such results were first reported by *Schmidlin*, [1976]. Later *Chandra* [1980] explained this by using a one-dimensional model to demonstrate that such features might be a direct result of turbulent mixing and dissipation. *Hauchecorne et al.* [1987] have made a statistical study and with crude estimation concluded that the gravity waves in the presence of an inversion layer would break (due to the mixing and viscous dissipation associated with turbulence), preferably, inside and above the layer. Figure 14 shows a schematic description regarding what happens during the mesospheric temperature inversion. According to the author, when the wave reaches the inversion layer, N (Brunt-Visalia frequency) increases suddenly and in the zonal wind above the westerly mesospheric jet and hence a super saturation will occur. This wave deposits its momentum, decelerates the zonal wind and its breaking produces turbulence. This turbulence occurs inside and above the inversion layer. Details of turbulence produced with this breaking of gravity waves will be discussed in next the section. The turbulence produced above the minimum temperature helps in maintaining the inversion layer by two processes [*Hauchecorne et al.* [1987]. The first one is the heating due to the viscous dissipation of turbulent motions and the second one is the downward vertical heat flux from the upper layer to the inversion layer due to turbulence mixing of the atmosphere.

A similar kind of mesospheric inversion around 70 km was observed by *Whiteway et al.* [1995(a)] at Toronto (44^0 N, 80^0 W). They found that the percentage of occurrence of such inversions during different seasons had a good correspondence with the MST radar echoes and climatology of gravity wave activity. The percentage occurrence of Indian MST radar echoes and Nd:YAG lidar temperature inversions over Gadanki (13.5^0N, 79.2^0E) are shown in Figure 15. From the figure it is clear that the percentage occurrence of mesospheric echoes and temperature inversions have a good correlation. These echoes at mesospheric heights occur in the form of highly reflecting layers (see Figure 8 for more details) as reported by several investigators [*Rottger*, 1987; *Yamamoto et al.*, 1987; *Murakao et al.*, 1988, *Tsuda et al.,* 1990]. In these studies, the observed echoing layers have been discussed on the basis of turbulent scattering, and the enhanced echoes have been related to the turbulence generated by the breaking of tides and gravity waves. This aspect is well documented by *Fritts* [1989] and references therein.

Thomas et al., [1996] also observed the mesospheric structures both above and below the temperature inversion using the coordinated MST radar and Lidar data. *Namboothiri et al.,* [1996], using simultaneous observations of MU radar (in meteor mode) and Sodium lidar observed the same gravity waves. Thus the observed high reflectivity in the 74-76 km region could be due to a well-mixed turbulent layer as reported by different investigators [*Haunchecrone et al.*, 1987; *Whiteway et al.*, 1995 (b); *Thomas et al.*, 1996].

Turbulence studies:

Introduction:

A study of turbulence in the free atmosphere and its relationship to gravity wave and tidal breakdown can be more effectively studied using this technique. In the mesosphere and lower thermosphere, turbulent echoes are only observable during daylight (i.e., only when there is sufficient ionization and associated ionization gradients). Estimates of turbulence intensity and turbulent diffusion appear to be possible using high spatial resolution radars with narrow antenna beams to study the space-time structure of atmospheric turbulence layers.

The mechanism of turbulent scatter which was invoked by *Pekeris* [1947], *Booker* and *Gordon* [1950] and later developed by *Tatarskii* [1971] to explain over-the-horizon tropospheric radio propagation and this has been applied successfully to explain VHF and UHF radar echoes from the clear atmosphere [*Hardy et al.*, 1966; *Krofli et al.*, 1968; *VanZandt et al.*, 1978; *Green et al.*, 1979; *Watkins* and *Wand*, 1981].

The eddy dissipation rate has been measured by both *in situ* and remote sensing techniques by many groups [*Lilly et al.*, 1974; *Barat* and *Bertin*, 1984; *Lubken et al.*, 1987 with *in situ* techniques; *Frisch* and *Clifford*, 1974; *Crane*, 1980; *Gage et al.*, 1980; *Weinstock*, 1981; *Sato and Woodman*, 1982; *Woodman* and *Rastogi*, 1984; Fukao et al., 1986] with remote sensing techniques and are reviewed by *Hocking*, [1985, 1996, 1997]. Attempts have been made to compare ε derived from *in situ* measurements with that derived from radar observations [*Cohn*, 1995; *Bertin et al.*, 1997; *Delage et al.*, 1997; *Rao et al.*, 1997]. They found a fairly good correlation between these two techniques. *Fukao et al.* [1994] studied the seasonal variation of vertical eddy diffusivity, closely related to eddy dissipation rate, in troposphere, stratosphere and mesosphere using three years of data collected with the MU radar. All these studies except *Fukao et al.,* [1994] are of using only limited data sets at mid and higher latitudes.

Knowledge of vertical eddy diffusivity, K, is very essential because it is one of the important parameters in atmospheric studies. Various techniques have been evolved inorder to study the eddy diffusivity, which include rocket [*Lubken et al.,* 1987], air-craft [*Lilly et al.,* 1974] observations. Studies of *Woodman and Guillen* [1974] triggered the evolution of a new generation of VHF/UHF radars and are extensively used for the characterization of turbulence because of their high temporal and spatial resolutions. Though the theory for deriving K from radar measurements has been developed long back [*Sato and Woodman*, 1982], but the follow up implementation is slow due to various reasons [*Hocking*, 1997]. Most of these reasons are solved and reviewed by *Hocking* [1983, 1985, 1986, 1997].

There are two methods proposed for the estimation of K from the radar measurements. All these methods assume that the turbulence is isotropic and in the inertial sub-range. Further, it is also assumed that the spectrum follows a kolmogoroff shape and the atmosphere is stably stratified ($N^2 > 0$). First method uses the radar-backscattered power for the quantification of turbulence; however, this method requires additional measurements of temperature and humidity with high resolution. The second method uses the spectral width of the backscattered echo for the estimation of K. This method is used in the present study and is described in detail in section 2. These two methods are extensively compared [*Cohn*, 1995; *Rao et al.,* 1997] and a reasonably good agreement has been found between the two methods. However, mainly two non-turbulent effects, i.e., beam broadening and shear broadening, contaminate the observed spectral width. Beam broadening effects will be only for vertical pointing beam and shear broadening will occur in all the beams. Before going to the actual calculation of eddy diffusivity using the width method, the effects due to non- turbulent parameters should be removed. The theory to remove these effects has been well developed and discussed extensively by *Atlas et al.* [1969], *Frisch and Clifford* [1974], *Sato and Woodman* [1982], *Hocking*, [1983,1985,1986], *Fukao et al.,* [1994], *Nastrom* [1997] and it is briefly discussed here.

Methodology for the estimation of Eddy diffusivity:

The observed radar spectral width is mainly contaminated by beam broadening and shear broadening effects. *Atlas* [1969] discussed these effects and presented some formulae, but they were applicable only for narrow beams. Later, generalized formulae are presented and are used by several investigators [for example *Hocking,* 1983, 85, 88 and 97; *Fukao et al.,* 1994; *Nastrom and Eaton*, 1997].

The spectral broadening due to finite beam width is given by [*Fukao et al.*, 1994]

$$\sigma_{beam} \approx \delta_{1/2} | \bar{u} | \qquad \text{---------- (1)}$$

where $\delta_{1/2}$ is the half-power half width of the effective radar beam and $\bar{u}$ is the horizontal velocity.

The effect of spectral broadening due to horizontal shear width is given by

$$\sigma_{shear} = \frac{1}{2}\left|\frac{\partial \bar{u}}{\partial z}\right| \Delta z Sin 10^0 \qquad \text{---------- (2)}$$

for a beam with a zenith angle of 10^0. where Δz is range resolution.

Besides these two effects one more namely transience effect, due to incoherent integration, also contaminates the observed spectral width and the contribution of this term will also be there to the observed width.

$$\sigma^2_{trans} = \overline{4(u^1_{2\tau})^2} \qquad \text{---------(3)}$$

where τ is the length of time used for incoherent integration.

After substracting these effects with the observed width, we get the width due to turbulence alone given as,

$$(\sigma_{turb})^2 = (\sigma_{obs})^2 - (\sigma_{beam})^2 - (\sigma_{shear})^2 - (\sigma_{trans})^2 \qquad \text{------------ (3)}$$

From this width (after correction), the eddy diffusivity can be calculated using the following equation

$$K \approx 0.1\frac{\sigma^2}{N} \qquad \text{---------- (4)}$$

Where N is the Brunt-vaisala frequency and is estimated from temperature measurements.

Most of the studies on eddy diffusivity are based on observations of short duration, except a few climatological studies. Moreover, these studies are confined to mid and high latitudes. *Fukao et al.,* [1994] studied seasonal variation of vertical eddy diffusivity, in the MST region using three years of data collected with MU radar. *Nastrom and Eaton* [1997] studied the climatology of eddy diffusivity using the VHF radar observations at WSMR, New Mexico. *Hocking* [1988] measured the turbulence parameters in the upper mesosphere and lower thermosphere with 2-MHz narrow beam radar for about two years over Adelaide, Australia. In tropics, climatological studies of turbulence parameters are not available. Moreover, the knowledge of K is necessary at various geographical locations inorder to understand the global circulation.

Some Results:

Figure 16 shows seasonal variability of the monthly medians of vertical eddy diffusivity K observed by the MU radar in the mesosphere [after *Fukao et al.,* 1994]. It is generally observed that the values of K are larger in the mesosphere than lower atmosphere and gradually increase with height. The seasonal maximum of K in the mesosphere over MU radar is observed in summer. It is also noticed that the gravity wave activity and the resultant momentum flux and energy density observed by the MU radar or by lidars in the mesosphere is maximum in summer. Similarly, seasonal variations of vertical eddy diffusivity over Gadanki (a tropical latitude) are shown in Figure 17. From the figure it is clear that during the summer and equinoxes, the maximum of K is observed. In the mesosphere, echoes will occur in the form of reflecting layer structures at these heights and can be interpreted them to be due to turbulence created by breaking of gravity waves. It is now generally accepted that dynamics and structure of the mesosphere are strongly influenced by the propagation of gravity waves into this region.

Gravity waves originating in the troposphere can propagate into higher altitudes with increase in amplitude. A vertically propagating wave begins to break at the level where there is a sudden change in temperature lapse rate. At zones of gravity waves breaking, a large amount of energy and momentum are lost by the waves, which is already discussed in last section. The overall values reported so far are depicted in Figure 18 [after *Fukao et al.*, 1994]. The values reported over MU radar and Indian MST radar are very less when compared with the models and the possible reasons for this are explained in detail by *Fukao et al.*, 1994.

Some special features observed in the Mesosphere

Polar Mesospheric Summer Echoes (PMSE)

Thought the occurrences of mesospheric echoes are highly intermittent in space and time at all the latitudes, it has some special characteristics in different seasons at different latitudes. To mention a few Polar Mesospheric Summer echoes (PMSE) observed in polar latitudes and pre-sunrise echoes at mesospheric heights etc.,. More extensive work was carried regarding the PMSE and many review papers have come through the course of work. Among them the review paper by *Cho and Kelley*, [1993] and very recently *Cho and Roettger,*[1997] and the references therein covered a detailed history of their evaluation and their characteristics. To mention briefly, these echoes will occur in the altitude region of 80-85 km in the northern latitudes. The scattering mechanism related to these kinds of echoes is still ambiguous whether these are due to specular reflection or due to turbulence. Literature also says that these are also associated with noctilucent clouds. These echoes are mainly caused by inhomogeneities in the electron density with an order of half radar wavelength. The electrons in the extreme cold summer mesosphere are electrically coupled with large charged particles like water cluster ions, aerosols or ice particles and the electron diffusivity is markedly reduced resulting in the maintenance of inhomogeneities of electron density at the radar Bragg scale [*Cho and Kelley*, 1993]. Until recently it was thought that the presence of PMSE is a characteristic of only polar latitudes and the available literature also says that it is not only a polar latitude phenomenon but can also seen at mid-latitudes.

Pre-Sunrise mesospheric echoes

Another important observation at mesospheric heights is regarding pre-sunrise mesospheric echoes. The most important effect of solar ultraviolet radiation on the upper atmosphere is ionization of its constituent gases. The ionization commences from a height of about 60 km and extends upto the highest limits of the atmosphere [*Rottger*, 1982]. Due to the ionization of molecules such as NO and O_2, free electrons are produced. These free electrons play a vital role in creating refractive index inhomogeneities at mesospheric heights and are considered to behave as passive tracers of neutral gas [*Fukao et al.*, 1980; *Hocking*, 1985]. These free electrons are responsible to form the D-region of the ionosphere in daytime. This region is mainly an absorbing region and disappears during nighttimes due to recombination with ambient molecular ions. Radar probing for lower ionospheric regions has started long back and it is well established that echo power above the noise level is detected only in daytime [*Gage and Balsley*, 1984]. Strong echoes at mesospheric heights can be detected under two conditions. One is the presence of free electron density gradients and the other is the presence of strong atmospheric turbulence. It is not possible to detect echoes even if one of them is lacking [*Muraoka et al.*, 1998]. But under some circumstances, it is possible to detect mesospheric echoes even during night-times, especially, just before sunrise and after sunset for some time. Figure 19 shows the echo power measured in the vertical beam by MU radar between 0330 and 0630 LT on 6 June 1995 [After *Muraoka et al.*, 1998]. This was recognized long back using different techniques [*Subbaraya et al.*, 1985] and explanations were given satisfactorily for appearance of this kind of strong echoes during pre-sunrise and post-sunset hours. According to *Mitra* [1992], ionization increases when ground – grazing rays strike the layer and the electrons produced cannot be due to ionization, but the increase may be due to electron detachment from the negative ions, which are known to have quite low detachment potentials.

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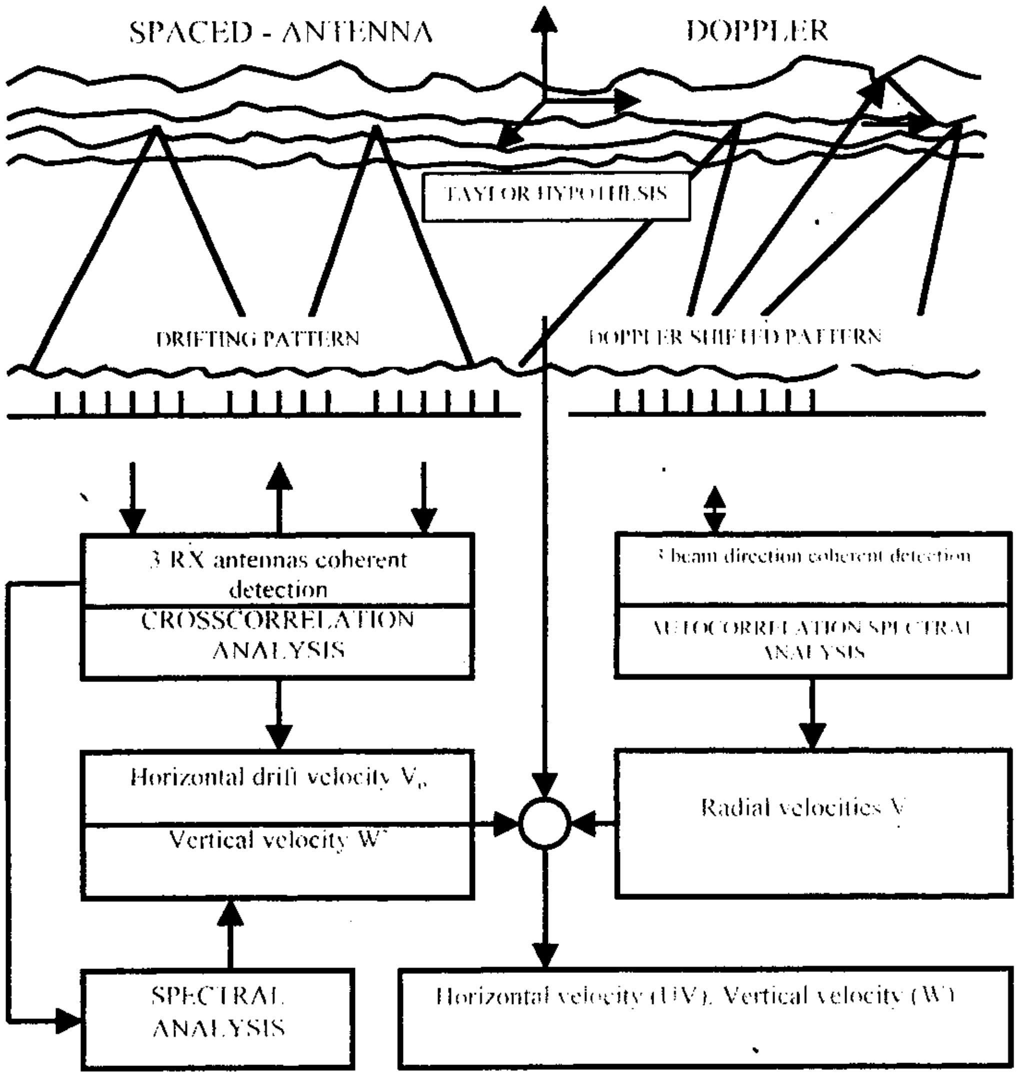
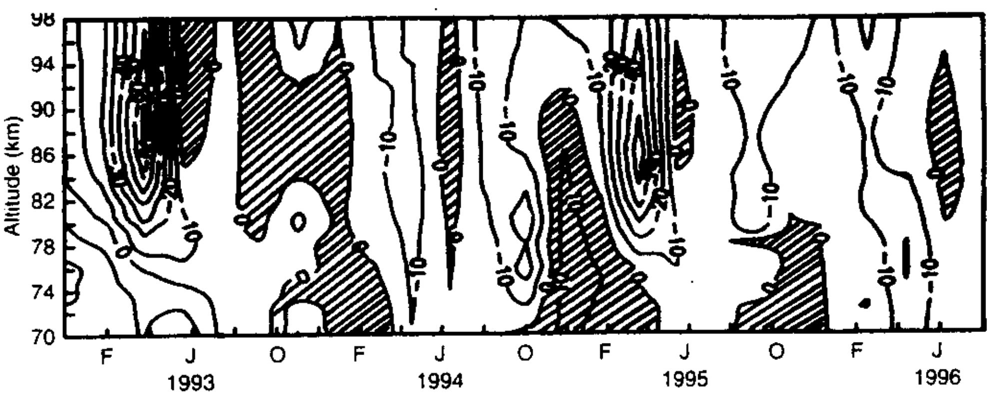


A schematic representation of the Doppler Beam Swinging (DBS) and Spaced Antenna (SA) methods of drift velocity measurements [Rottger, 1981]

Figure. 1

Mar. – Apr. – Spring Equinox
Sep. – Oct. – Fall Equinox
Nov. – Feb. – Winter
May – Aug. – Summer

(After Rajaram and Gurubaran, Ann. Geophysica, 16, 197-204, 1998



Mean (30-day averaged) zonal wind over Tirunelveli from Dec. 1992 to Aug. 1996. Contours provided with slanted lines represent regimes of eastward flow

Westward flow during Equinoxes

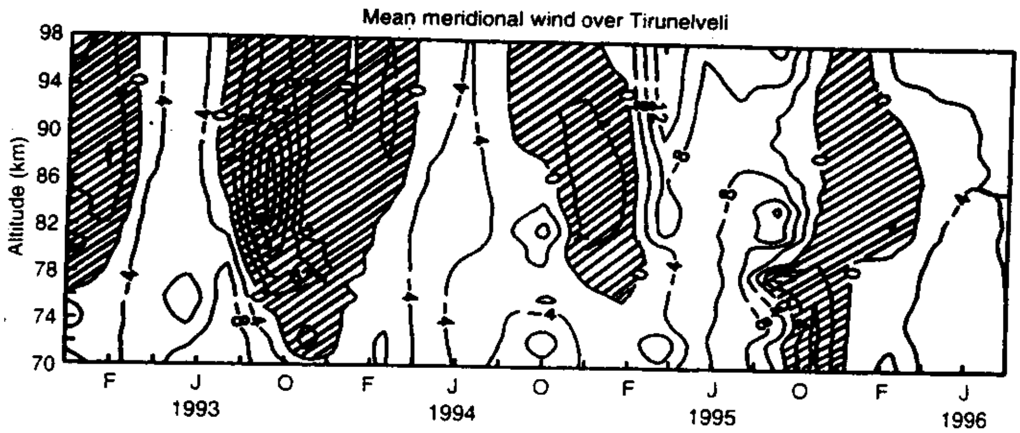
Eastward floe during Solstices

Downward trend on many occasions

Peak westward flow during 1993 & 1995 which is due to QBO

Figure. 2

(After Rajaram and Gurubaran, Ann. Geophysica, 16, 197-204, 1998)



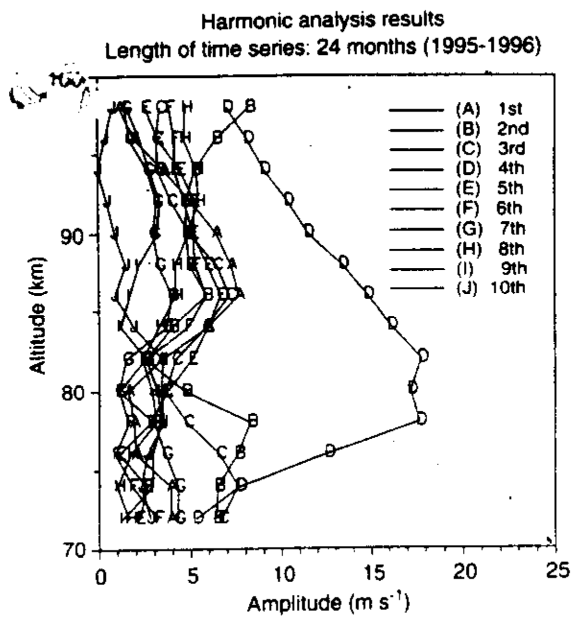
Mean meridional wind. Contours provided with slanted lines represent regimes of poleward flow

Poleward motion during Winter

Equatorward motion during Summer

Figure. 3

(After Rajaram and Gurubaran, Ann. Geophysica, 16, 197-204, 1998)



Harmonic analysis results for the mean zonal over a period of two years.

<u>Dominant motions in Equatorial Mesosphere</u>

A- QBO- Below 82 km – low amplitudes (3-4 ms^{-1})
Above 86 km - ~8 ms^{-1}

B- AO- Peak at lower heights

C- 8 months – Peak at 86 km

D- SAO- Peak around 80 km

Figure. 4

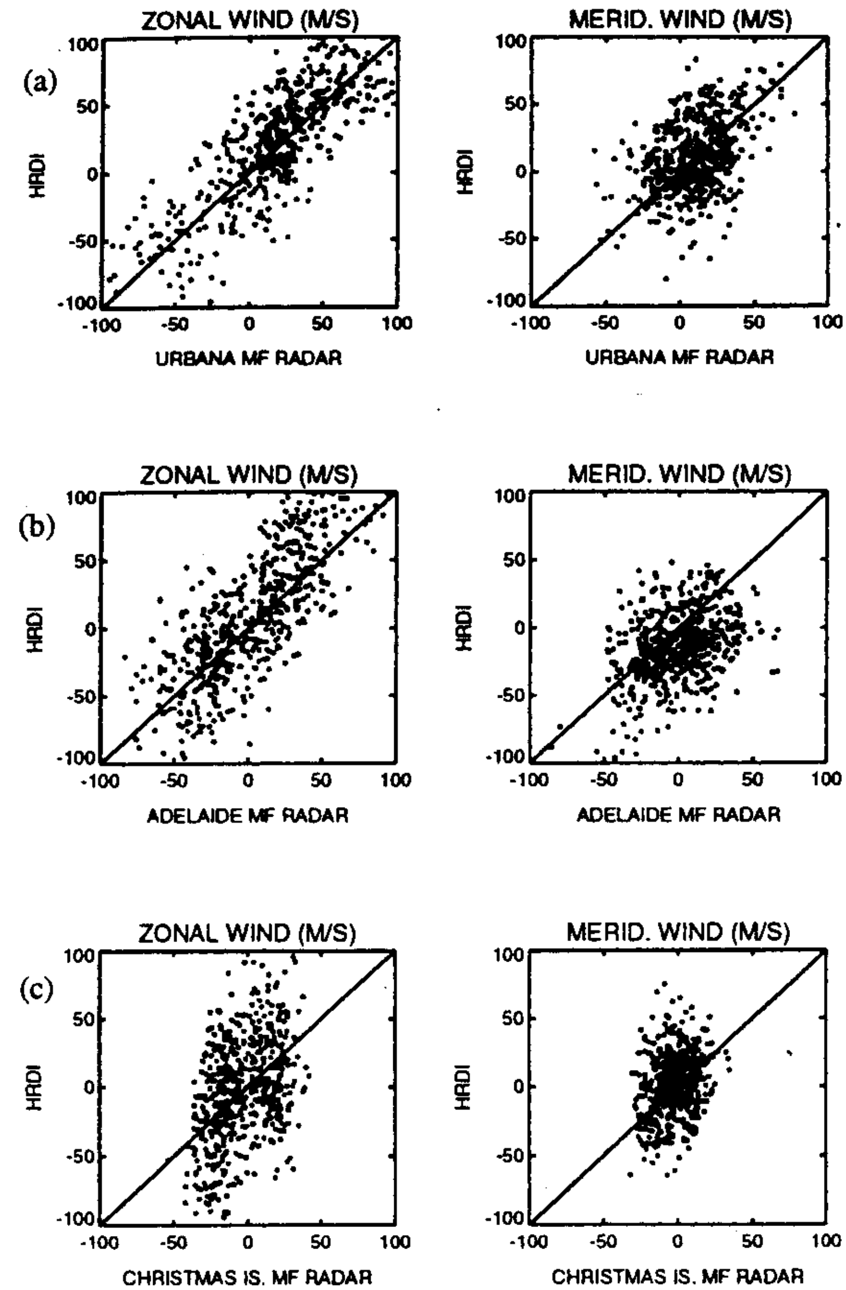
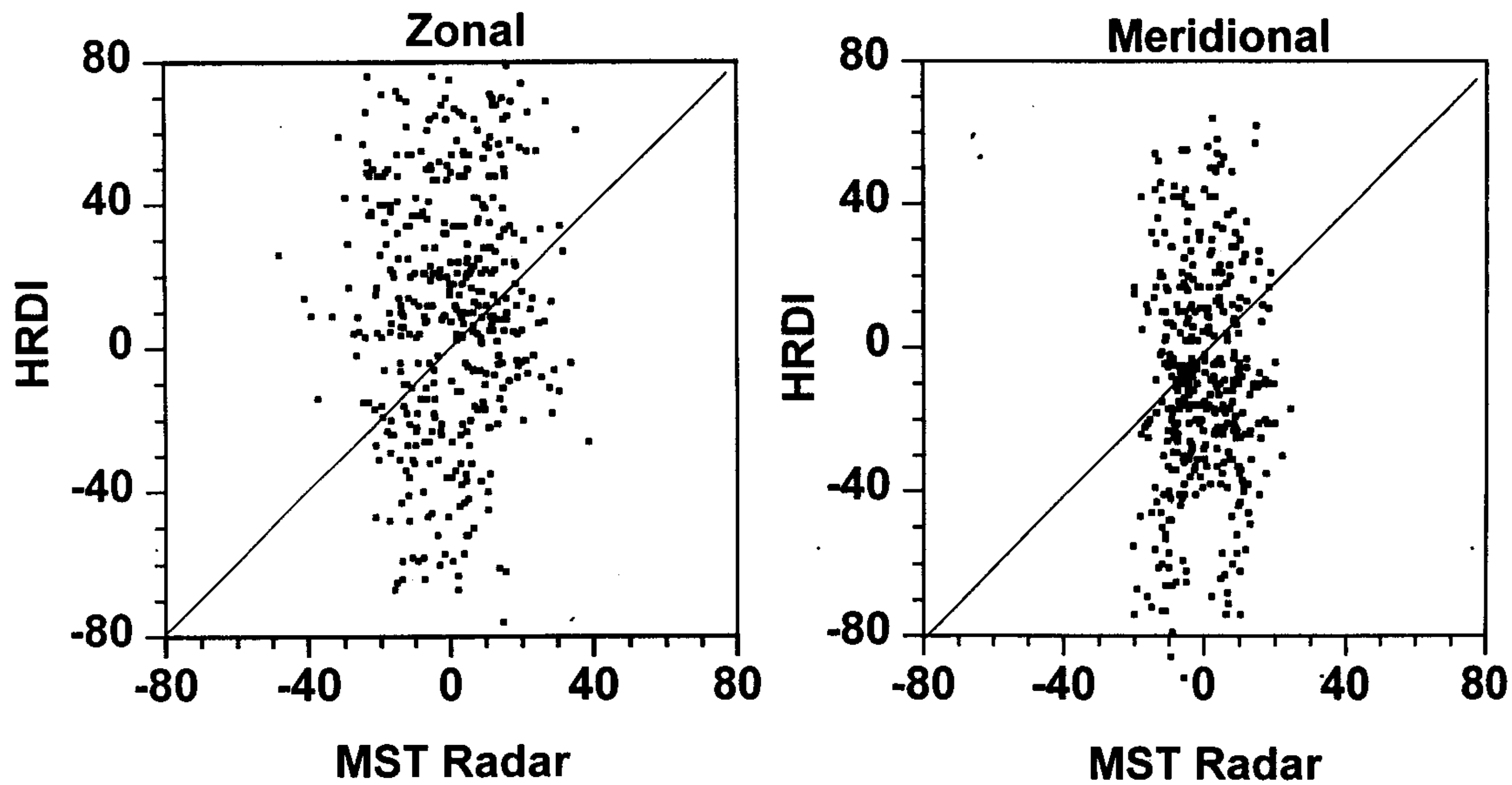


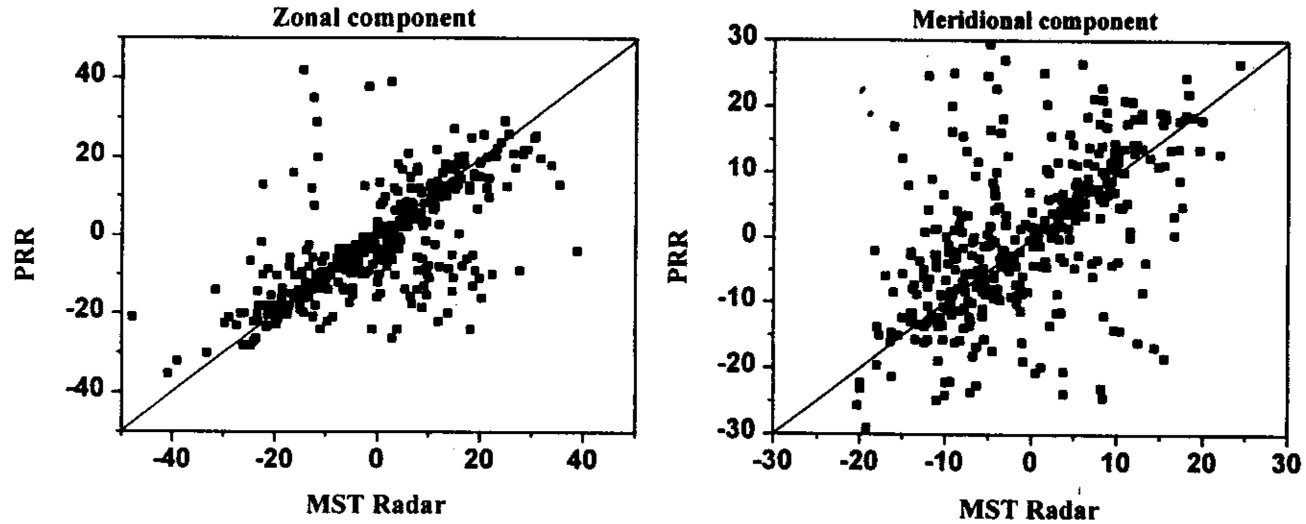
Figure . Scatterplots of HRDI winds in the altitude range 65-85 km for the zonal and the meridional components using (a) 106 coincidences with the Urbana MF radar between December 1991 and December 1993, (b) 118 coincidences with the Adelaide MF radar between December 1991 and January 1994, and (c) 137 coincidences with the Christmas Island MF radar between December 1991 and January 1994.

Figure. 5



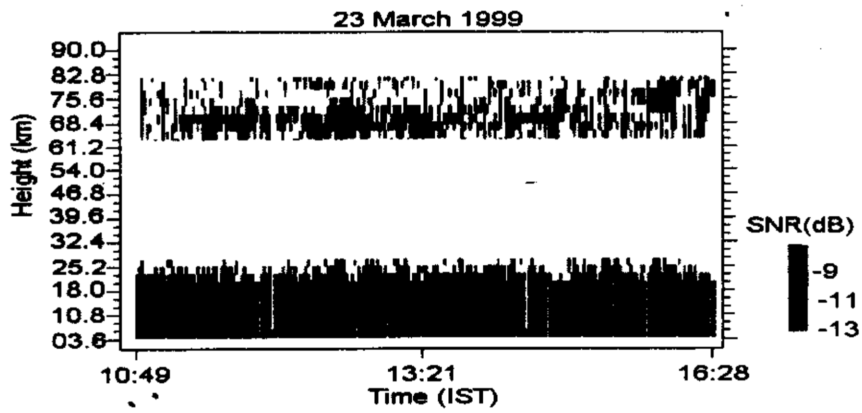
scatter plots between MST radar winds and UARS/HRDI winds in zonal and meridional components

Figure. 6



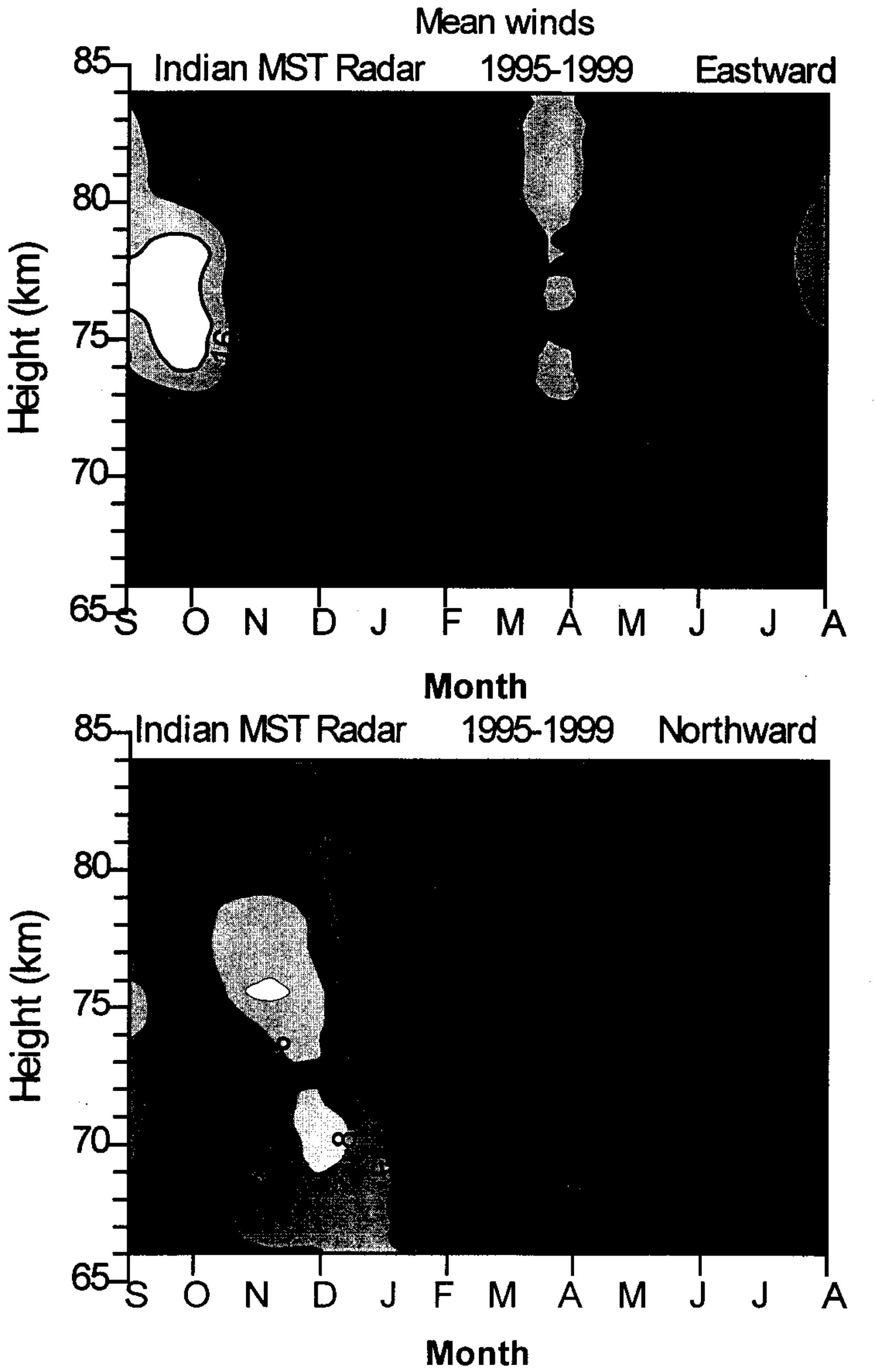
Scatter plots between MST radar winds and Partial Reflection Radar (PRR) winds in zonal (left) and meridional component (right).

Figure. 7



Indian MST radar observations of echo characteristics observed on 23 March 1999 in the vertical incidence

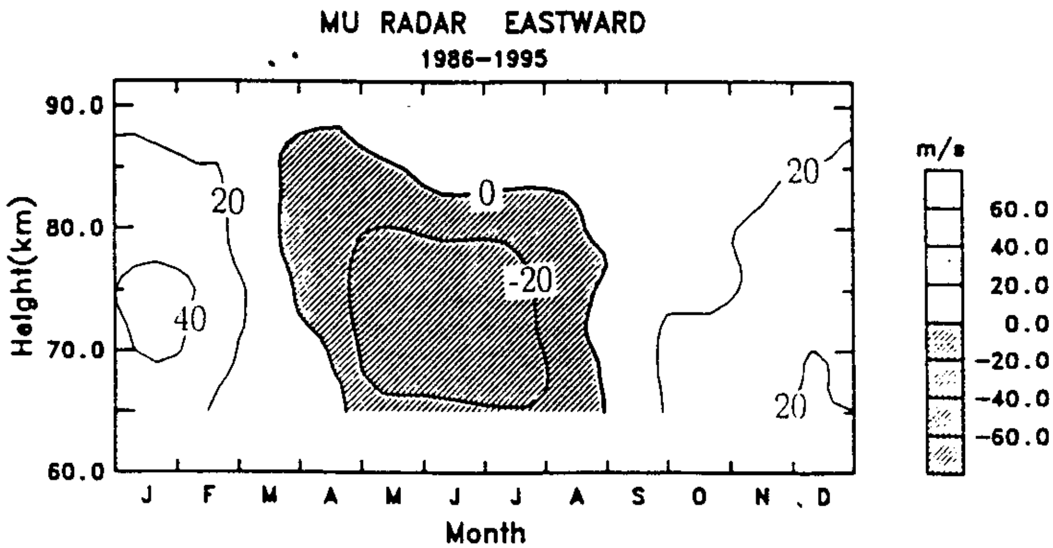
Figure. 8



mean zonal and meridional winds observed over Gadani using Indian MST radar

Figure. 9

Namboothiri et al., JASTP, 61, 1111-1122, 1999



Mean zonal winds averaged for the 10-year period. 1986-1995.

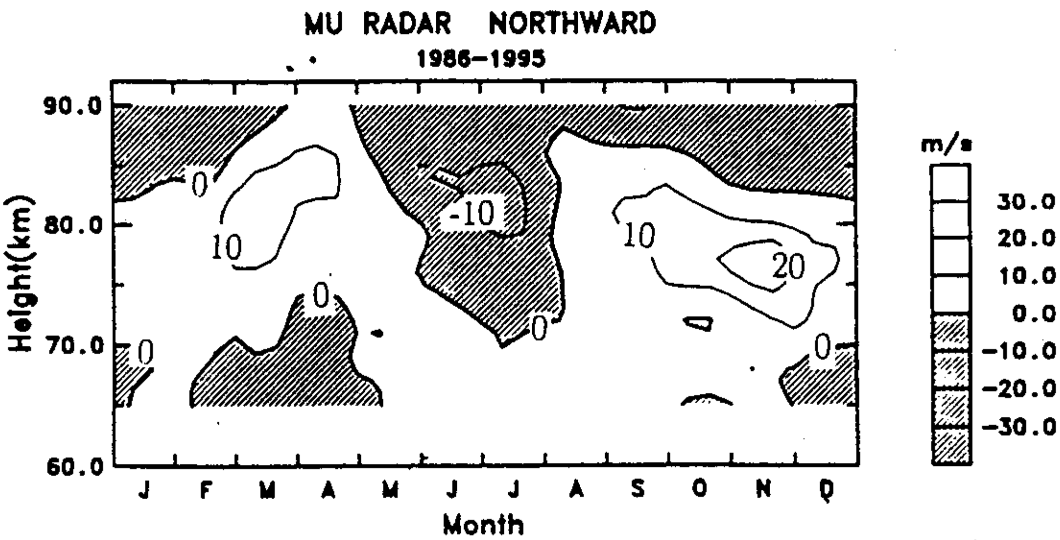
Westward flow during Summer

Eastward flow during rest of the seasons

AO is dominant mode

Figure. 10(a)

Namboothiri et al., JASTP, 61, 1111-1122, 1999



Mean meridional winds averaged for the 10-year period. 1986-1995.

Southward flow during Summer

Poleward flow during rest of the seasons

Figure. 10(b)

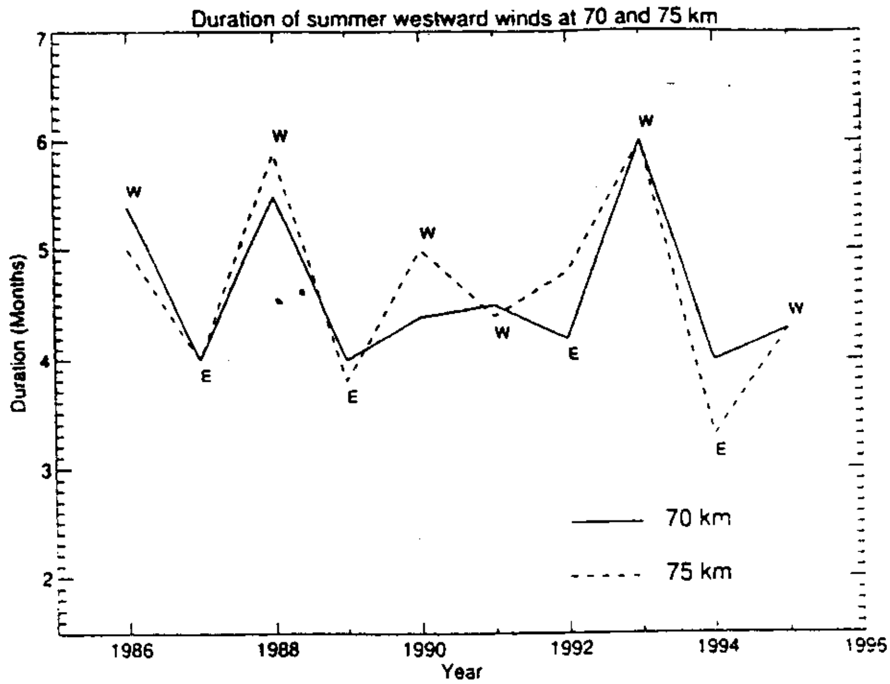


Fig. [illegible]. Duration of summer westward winds for the period 1986–1995. E and W represent the phases of equatorial QBO at 40–50 hPa level for each year.

Figure. 11

Charles and Jones, JASTP, 61, 351-362, 1999

Hally, Antartica

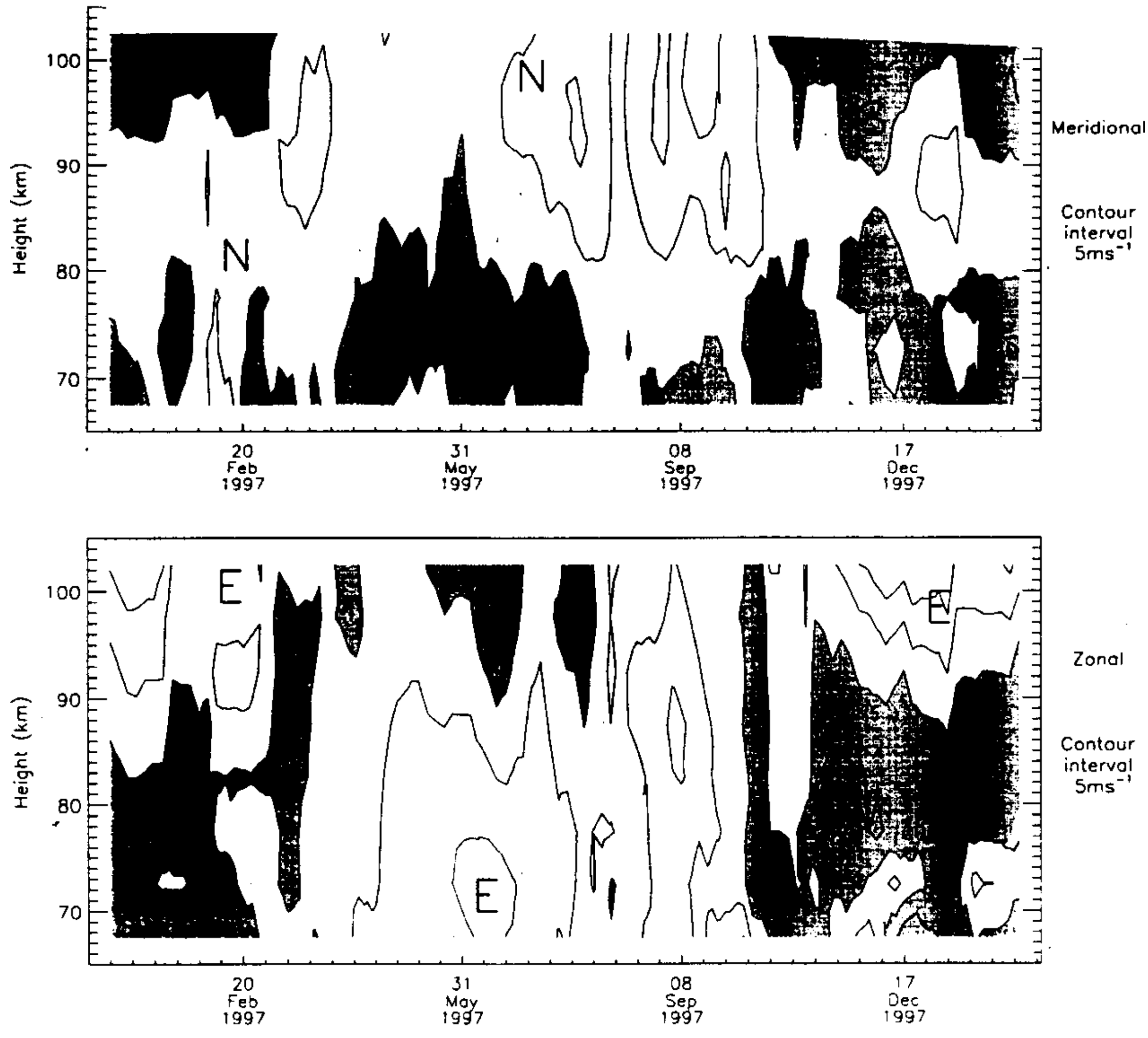
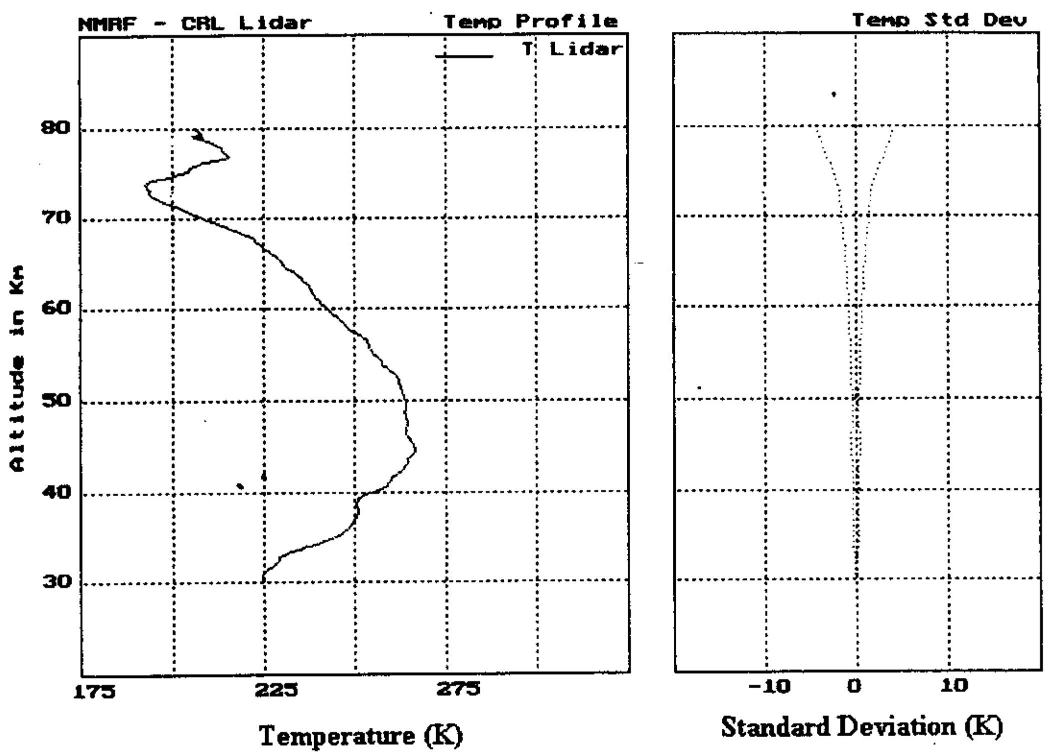


Fig. 2. Contour plot showing the seasonal variation in daily mean winds above Halley during 1997. The contour interval is 5 m s⁻¹ with southward and westward winds shaded.

Figure. 12



Typical temperature profile along with limits of uncertainity recoded on 8 May 1998

Figure. 13

Hauchecorne et al., GRL, 14, 933-936, 1987

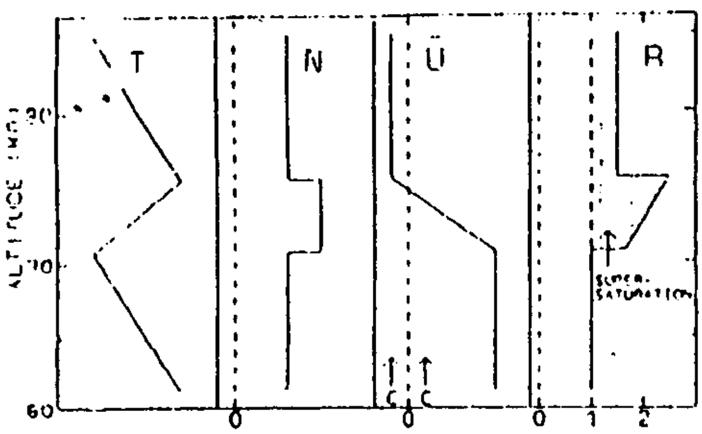
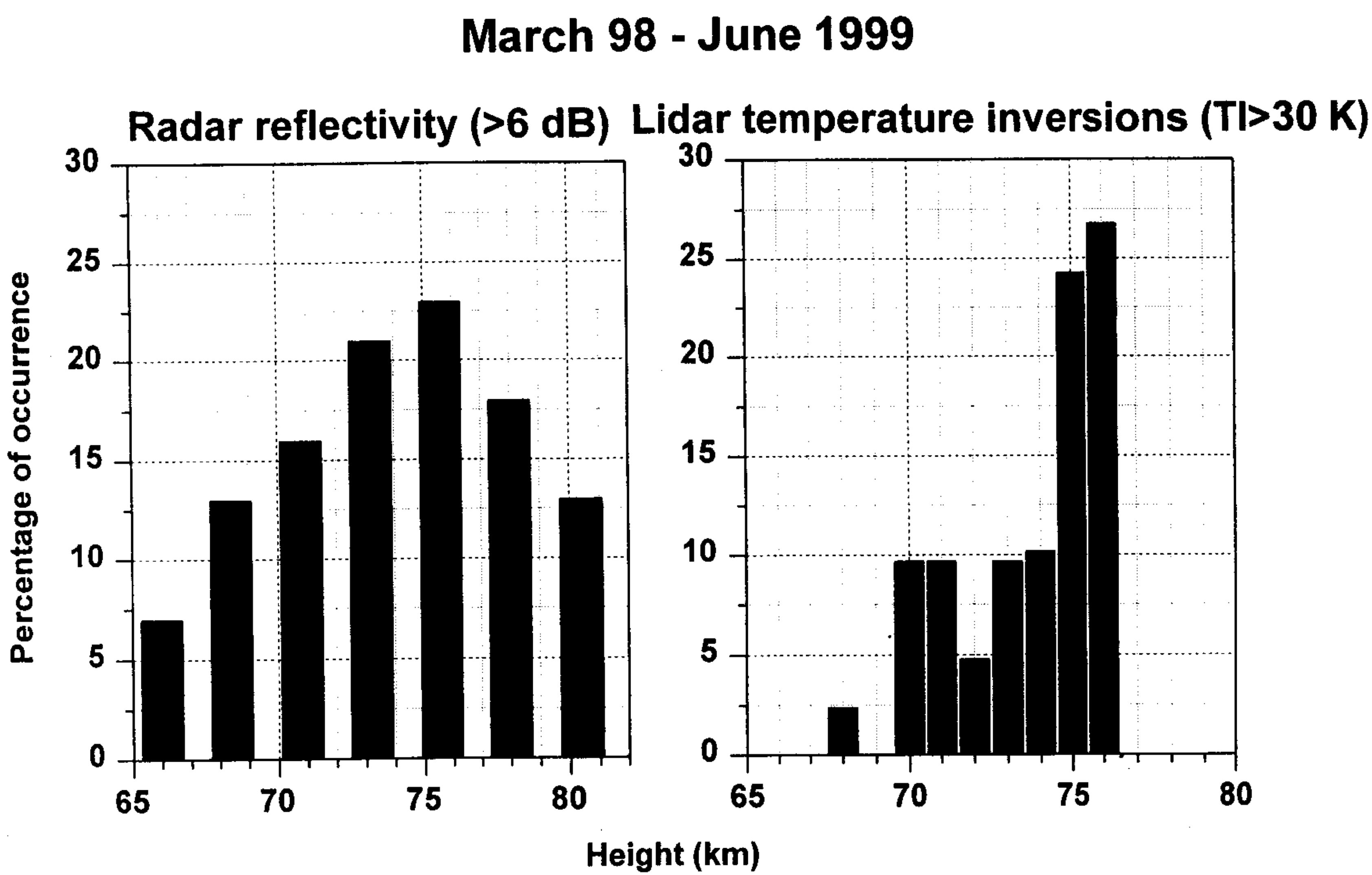


Fig. ` Schematic representation of the vertical profiles of the mean temperature T, the Brunt-Vaisala frequency N, the mean zonal wind $\bar{u}$ and the ratio of saturation R = T'/T'$_s$ during a mesospheric inversion.

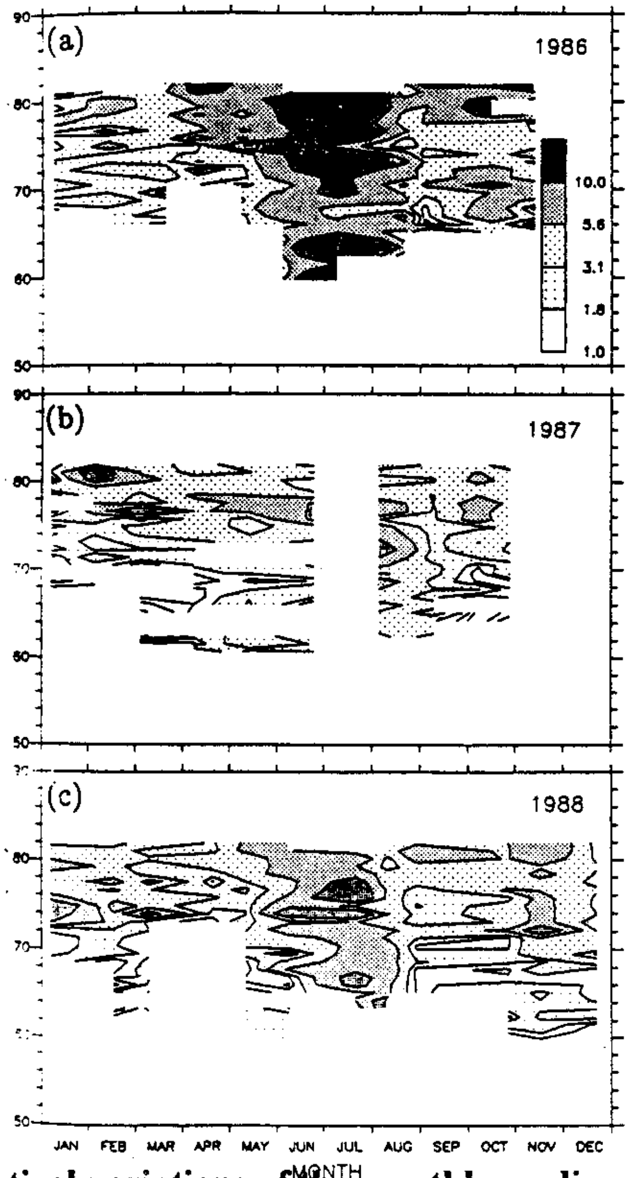
Figure. 14



Percentage occurrence of Indian MST radar echoes and Nd:Yag temperature inversions at different heights over Gadanki during March 98 to June 99

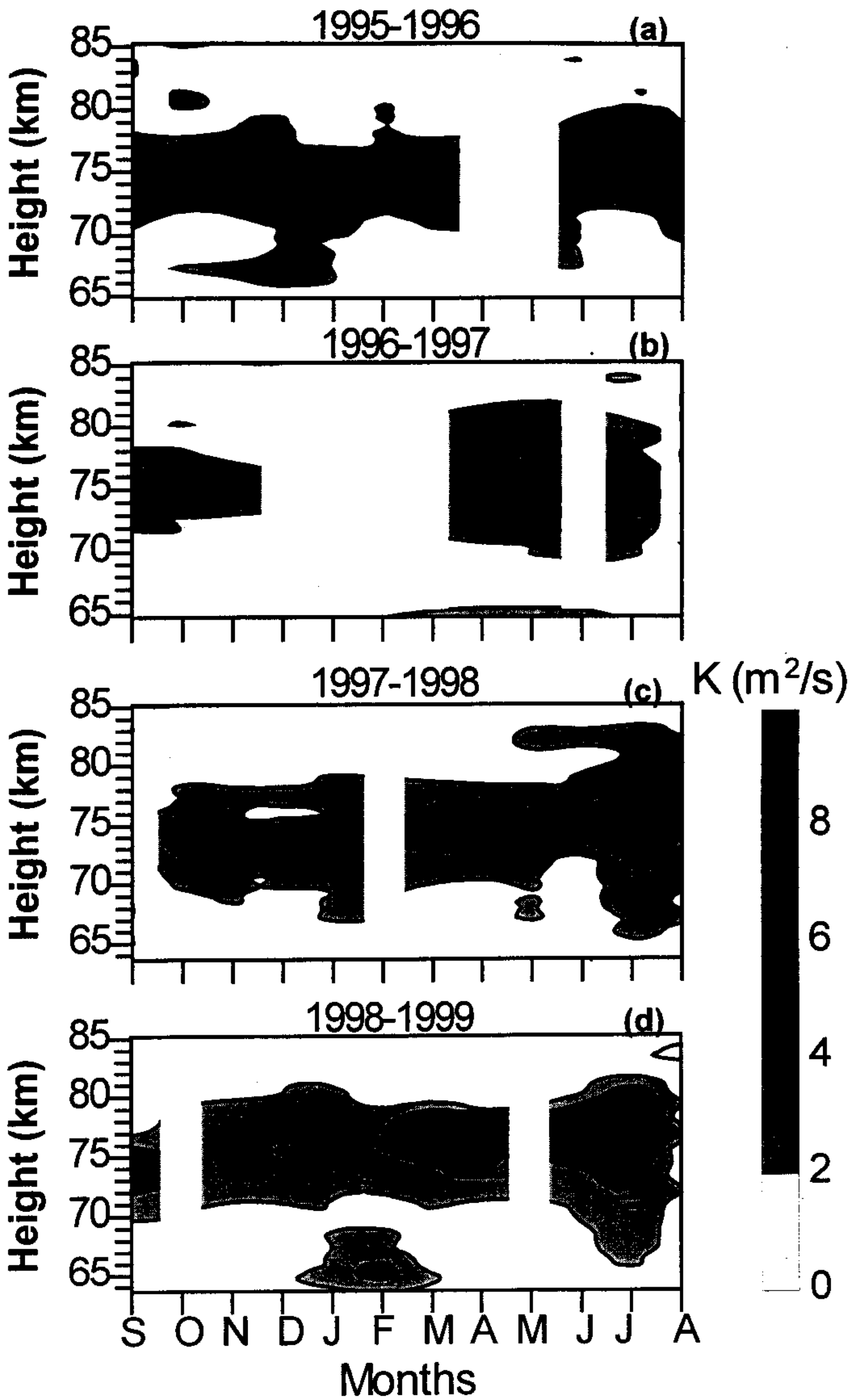
Figure. 15

Fukao et al., JGR, 99, 18,973-18,987, 1994



Seasonal vertical variations of the monthly medians of vertical eddy diffusivity K ($m^2 s^{-1}$) in the mesosphere over MU radar

Figure. 16



Monthly medians of vertical eddy diffusivity, K observed over Gadanki during 1995-1999

Figure. 17

Fukao et al., JGR, 99, 18,973-18,987, 1994

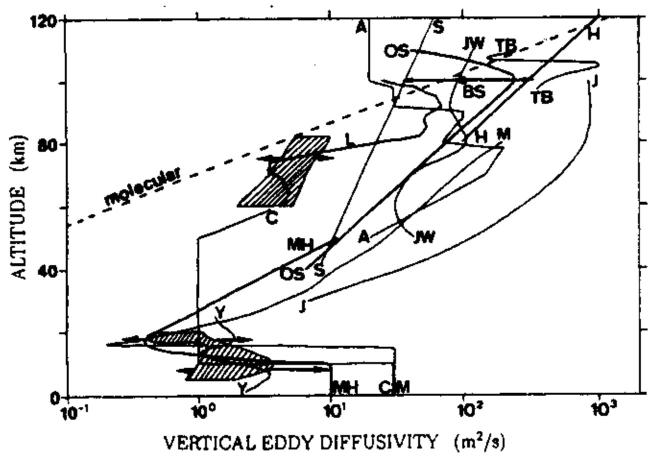


Figure ⁴. Comparison of the present observations (the variability of the annual medians is shown by the hatched areas) with previous studies (*A* [*Allen et al.*, 1981]; *BS* [*Blum and Schuchardt*, 1978]; *C* [*Crutzen*, 1974]; *H* [*Hocking*, 1985b]; *J* [*Justus*, 1973]; *JW* [*Johnson and Wilkins*, 1965]; *L* [*Lübken et al.*, 1993]; *M* [*McElroy et al.*, 1974]; *MH* [*Massie and Hunten*, 1981]; *OS* [*Ogawa and Shimazaki*, 1975]; *TB* [*Teitelbaum and Blamont*, 1977]; *S* [*Strobel et al.*, 1987]; and *Y* [*Yamazaki*, 1989]). The seasonal variabilities obtained by the present observations (as the ranges of the seasonal medians) and by the study of Blum and Schuchardt are shown by arrows. An approximate profile of the molecular diffusivity is also indicated by broken curve.

Figure. 18

Muraoka et al., 25, 2393-2396, 1998

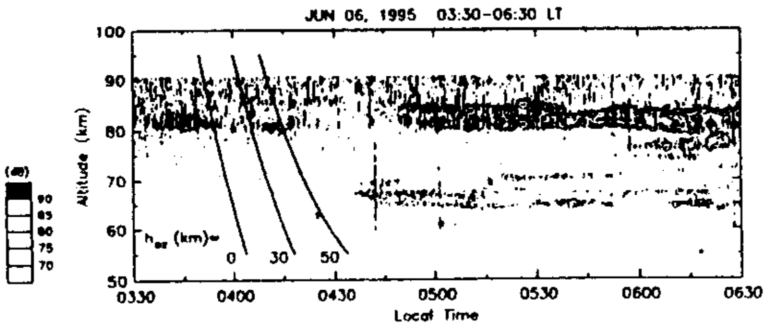


Figure ˜. Time-height contour of echo power measured in the vertical beam of the MU radar between 0330 and 0630 LT on June 6, 1995. The echo power is in an arbitrary unit and contours are drawn at 5 dB intervals. Three lines in the figure give the time of sunrise as a function of height in the respective cases where solar radiation is masked by the ozone layer located at the denoted altitudes (h_{oz}).

Figure. 19