 doi: 10.1256/qj.03.86

# Estimation of turbulence parameters in the lower atmosphere from MST radar observations

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(Received 29 May 2003; revised 14 November 2003)

SUMMARY

Using the Mesosphere-Stratosphere-Troposphere (MST) radar observations at Gadanki (13.5°N, 79.2°E) in the troposphere and lower stratosphere, the nature of the temporal spectrum of the vertical wind in the inertial sub-range is investigated. It is found that the slope of the spectrum depends upon the background horizontal wind speed. The turbulent-kinetic-energy dissipation rate $\varepsilon$ has been estimated using the MST radar observations by (i) the Doppler spectral-width method, (ii) the backscatter signal-power method, and (iii) the velocity-variance method. A detailed comparison of the $\varepsilon$ in the troposphere and lower stratosphere estimated by the three methods has been carried out. It is found that $\varepsilon$ by the variance method differs from that by the other two methods very significantly in the troposphere where the background horizontal wind is high. In the lower stratosphere with low background horizontal wind, $\varepsilon$ from the three methods is different in each case. It appears that turbulence in the regions of high background horizontal wind in the troposphere is anisotropic and the differences between the estimates of $\varepsilon$ by the variance method and the other two methods are attributed to the presence of anisotropic turbulence.

KEYWORDS: Anisotropic turbulence Turbulence spectrum Vertical wind

## 1. INTRODUCTION

The turbulent-kinetic-energy dissipation rate $\varepsilon$ and the eddy diffusion coefficient $K_m$ are two of the important parameters affecting the atmospheric dynamics and trace constituent distribution. With the advent of Mesosphere-Stratosphere-Troposphere (MST) radar for probing the lower stratosphere and the troposphere, it has become possible to estimate the turbulence parameters from wind data owing to the high degree of altitude and temporal resolutions that can be obtained by the radar (Hocking 1989). There are basically three methods of estimating $\varepsilon$ (and hence $K_m$) from radar observations. These are (i) the Doppler spectral-width method, (ii) the backscatter signal-power method, and (iii) the wind-variance method. The assumptions and approximations involved in these three methods have been discussed in the literature (e.g. Gage and Balsley 1978; Hocking 1989; Fukao *et al.* 1994; Nastrom and Eaton 1997; Delage *et al.* 1997). VanZandt *et al.* (2000) devised a dual wavelength radar technique for the estimation of $\varepsilon$ using the radar returned signal power. However, this technique can be adopted only if two radars operating at different wavelengths are available. Recently, the variance method has been adapted with a new approach to estimate the turbulence parameters (Satheesan and Krishna Murthy 2002). The chief merit of this approach, which involves estimation of the temporal spectrum of the vertical wind, is that it is entirely based on the vertical wind data from the radar and thus does not require any complementary experiment.

Detailed investigations on the temporal spectrum of the vertical wind and on estimates of $\varepsilon$ from the three methods mentioned above have been carried out using the MST radar data at the tropical station Gadanki. The results of this investigation are presented in this paper.

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## 2. DATA

The vertical and horizontal wind data of MST radar at Gadanki are used in the present study. A detailed description and specifications of the radar are given in Rao *et al.* (1995). The radar operates at a frequency of 53 MHz with a range resolution of 150 m. The vertical antenna beam width is $\sim 3^\circ$ (one way). The radar antenna beam can be oriented along any of the five directions, $\pm 20^\circ$ off zenith in the north, south, east and west directions in the vertical plane and zenith. The wind data from the radar were collected for four seasons, from 18 January to 5 March 1999 (winter 1999), from 21 February to 1 April 2000 (winter 2000), from 2 July to 13 August 2001 (summer 2001) and from 4 February to 20 March 2002 (winter 2002), on each day. The vertical wind data were recorded on each day continuously for 2 h from 2000 to 2200 h IST (1430 to 1630 UTC) at $\sim$40 s intervals. The vertical wind data collection was followed and preceded by 10 min of oblique beam operation of the radar to obtain the horizontal wind data.

## 3. TEMPORAL SPECTRUM OF THE VERTICAL WIND

The vertical wind data of 2 h duration sampled at 40 s intervals on each of the observation nights are subjected to Fourier transform analysis to obtain the frequency–amplitude spectrum. This is then converted into a frequency–power spectrum. Pre-whitening of the data is not found necessary as a few test cases yielded no significant differences between the spectrum with and without pre-whitening. Further, before subjecting the data to Fourier transform analysis, wild data points, if any, are removed. From the power spectrum, the Brunt–Väisälä frequency $N$ is identified using the criteria given by Revathy *et al.* (1996). The $N$ demarcates the power spectrum of the vertical wind between two distinct frequency ranges, namely, the buoyancy range (or the range in which the wave propagation/disturbances can exist) at frequencies less than $N$, and the inertial sub-range of turbulence at frequencies greater than $N$ (Weinstock 1978).

From the temporal spectra in log-log scale, the slope $S$ of the best-fit line in the frequency range between $N$ and the Nyquist frequency is obtained along with its standard error. The slope thus estimated represents the slope of the spectrum in the inertial sub-range. Theoretically, the slope of the one-dimensional (isotropic) turbulence spectrum in the inertial sub-range is $-5/3$. As only the vertical wind velocity spectra are considered in the present study, the one-dimensional case is relevant. The altitude profiles of the spectral slopes obtained, corresponding to 29 January and 26 February 1999, are shown in Figs. 1(a) and (b), respectively, along with the error bars. The spectral slope varies with altitude over a considerably wide range from about 0 to $-3$. The fluctuations appear to be quite marked on 29 January compared to 26 February 1999. The spectral slopes are further examined by grouping the data into the four seasons. The slopes for each season are then grouped into six altitude ranges, namely 3.75–5.85 km, 6.00–8.85 km, 9.00–11.85 km, 12.00–14.85 km, 15.00–17.85 km and 18.00–21.00 km. Histograms of distribution of slope values are drawn for each season. The histogram for winter 2000 is shown in Fig. 2. The continuous line curve in the figure is the Gaussian distribution fit. The $\chi^2$ test for goodness of fit revealed that the Gaussian fit to the observed distributions is generally in the confidence level range of 95% to 99%.

The mean $S$ is plotted against the corresponding altitude range for the four seasons as shown in Fig. 3. All the mean values of $S$ are negative in spite of the presence of a few positive values. Starting from a high value in the lower altitude range, $S$ increases with altitude reaching a maximum in the altitude ranges 12–15 km and 15–18 km for the winter seasons and summer season, respectively, above which it decreases.

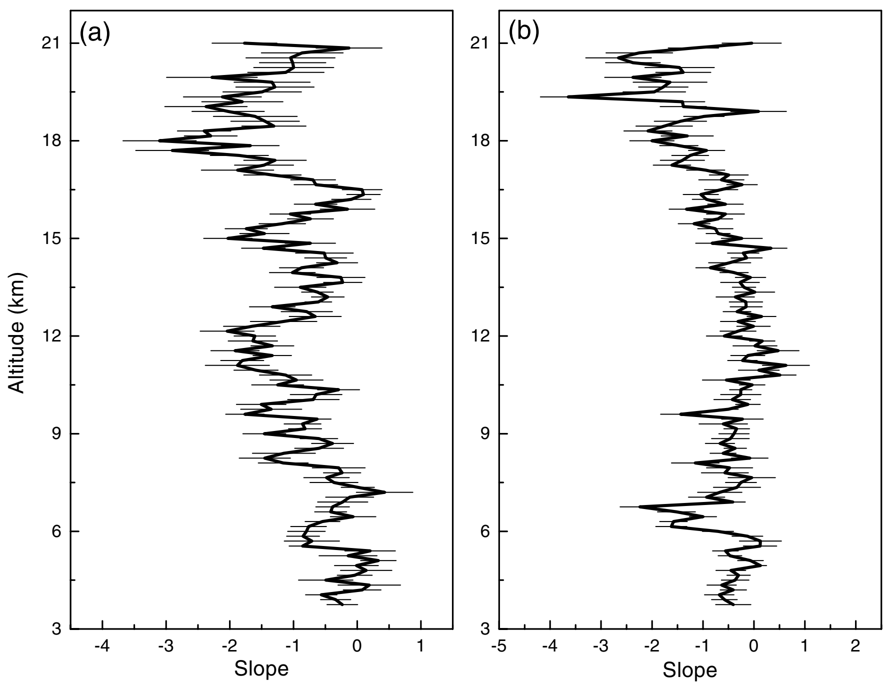


Figure 1. Altitude profile of *S* (see text) on 29 January and 26 February 1999. The horizontal bars are the error bars.

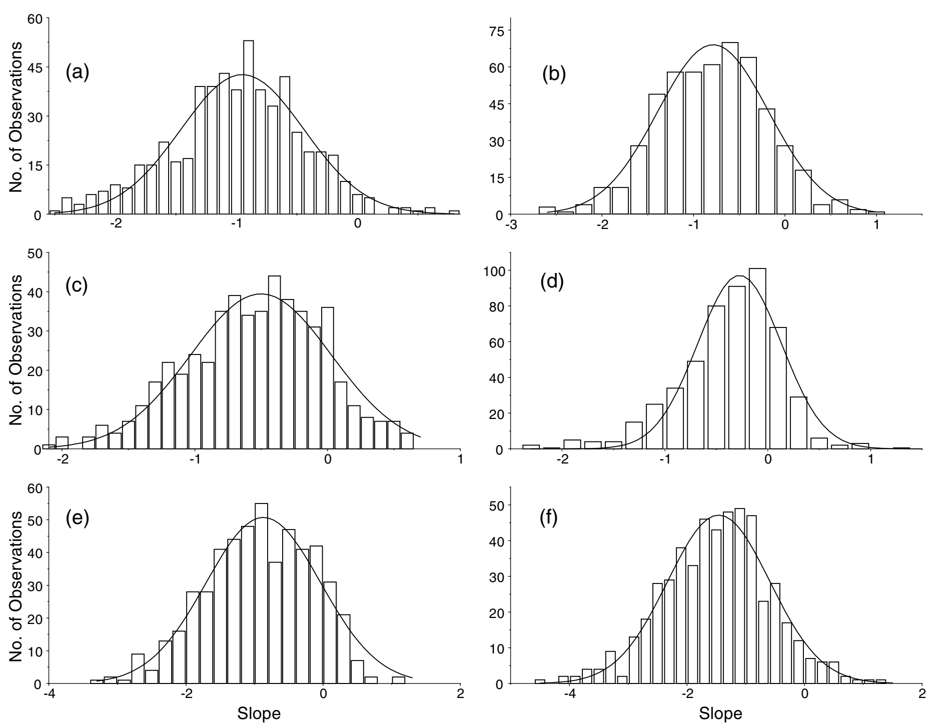


Figure 2. Histogram of the slopes for the altitude ranges (a) <6 km, (b) 6.00–8.85 km, (c) 9.00–11.85 km, (d) 12.00–14.85 km, (e) 15.00–17.85 km and (f) 18–21 km, for winter 2000.

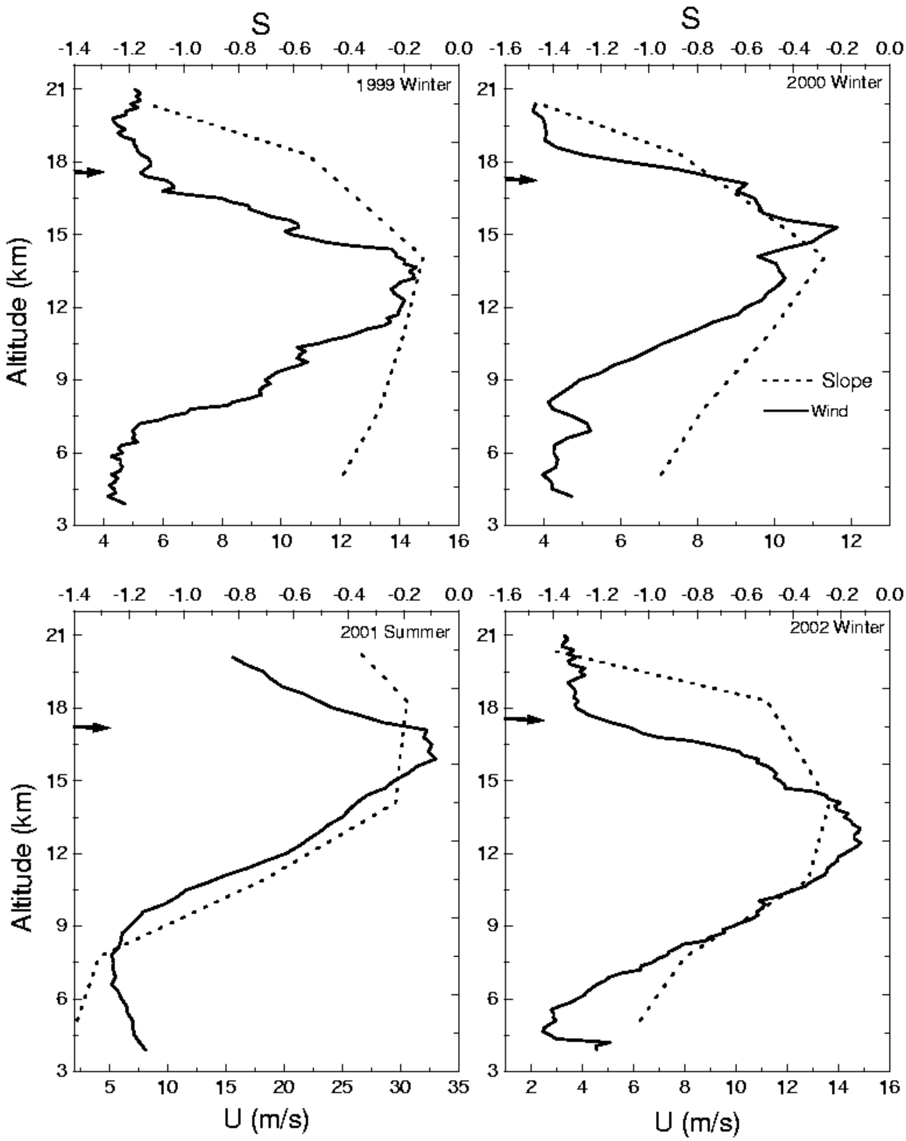


Figure 3. The altitude profiles of the mean horizontal wind $U$ (continuous line) and mean slope $S$ (dashed line) for the four seasons. The arrows mark the cold-point tropopause level (see text).

The maximum values of $S$ in summer and winter seasons are nearly the same. At the lowest altitude range $S$ is smaller in summer than that in winter, while at the highest altitude range (18–21 km, in the lower stratosphere), it is smaller in winter.

In order to study the relationship, if any, between $S$ and horizontal wind speed, the mean $S$ is plotted against the mean horizontal wind $U$ for the corresponding altitude for each season as shown in Fig. 4. The mean horizontal wind is obtained from the MST radar observations with oblique beams for 10 min preceding and 10 min following (total 20 min) the 2 h observation period with the vertical beam on each of the observation nights. The mean wind thus obtained can be taken to represent the mean over the 2 h period. It is seen that $S$ and $U$ are associated and that the functional form of the

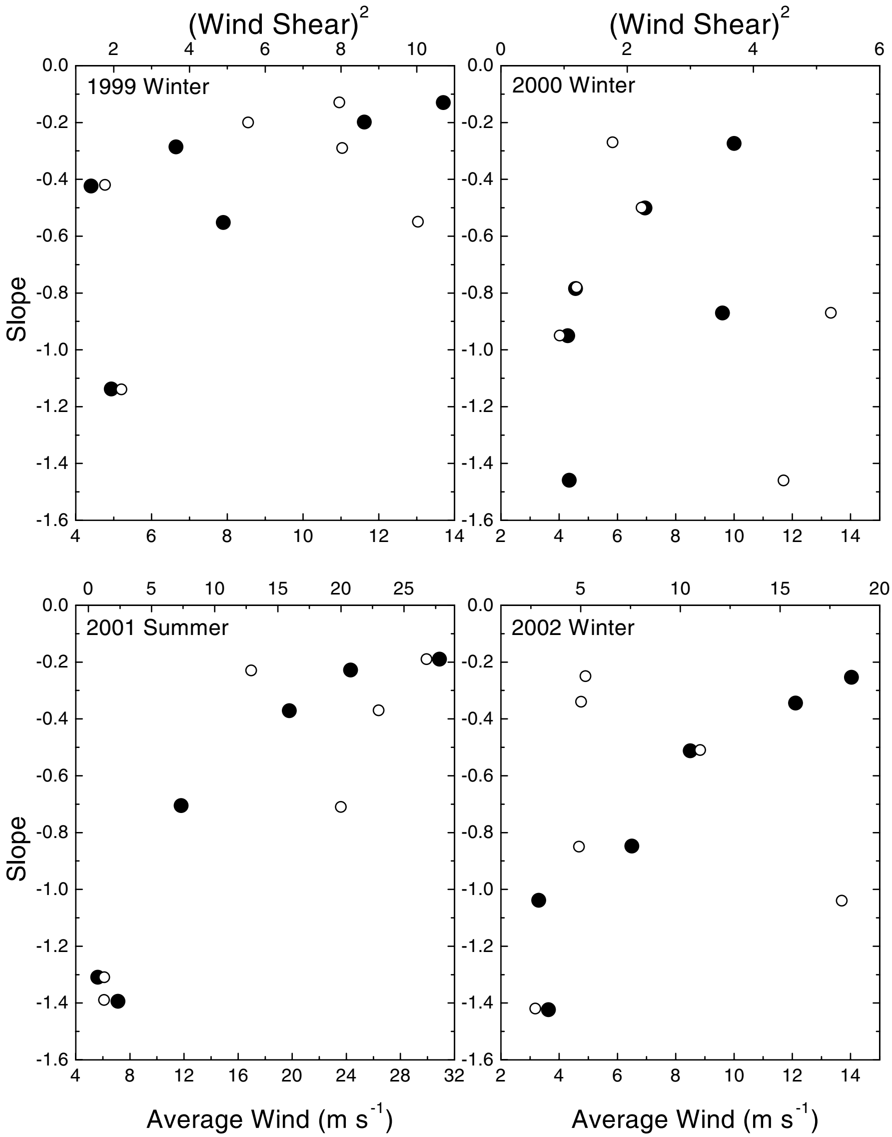


Figure 4. Slope vs. average horizontal wind (filled circles) and slope vs. square of wind shear (open circles) for winter 1999, winter 2000, summer 2001 and winter 2002.

relationship between the two is nonlinear. This is particularly evident for the summer season when the range of values of $U$ is high and also in winter 2002. An exponential function, given by

$$S = -A \exp(-BU), \tag{1}$$

is fitted to the data, and the values of $A$ and $B$ (along with the error) and the correlation (linear) coefficients between $\ln|S|$ and $U$ are given in Table 1.

The correlation coefficients are all significant at the probability $P < 0.05$, except the one for winter 2000 for which $P = 0.14$ (Fisher 1970). The values of the exponential coefficient $B$ are about the same for the three winter seasons, but $B$ is comparatively smaller for the summer season. This indicates that $S$ is more sensitive to the horizontal

TABLE 1. THE VALUES OF $A$ AND $B$ IN THE RELATION $S = -A\exp(-BU)$ ALONG WITH THE CORRELATION COEFFICIENT FOR FOUR DIFFERENT SEASONS

| Season | Correlation coefficient | A | B |
|---|---|---|---|
| 1999 winter | −0.82 | 0.35 ± 0.53 | 0.17 ± 0.06 |
| 2000 winter | −0.68 | 0.65 ± 0.58 | 0.15 ± 0.08 |
| 2001 summer | −0.98 | 0.76 ± 0.14 | 0.09 ± 0.008 |
| 2002 winter | −0.98 | 0.72 ± 0.14 | 0.15 ± 0.002 |

wind in winter than in summer. The values of $A$ are about the same for all the seasons except for winter 1999.

The $-5/3$ slope for the turbulence spectrum in the inertial sub-range is valid for isotropic (homogeneous) turbulence (Hocking 1983). In the atmosphere, however, large-scale disturbances would be anisotropic because of gravity stratification. In the vertical direction, the buoyancy scale would be limited by stratification, whereas in the horizontal direction no such restriction applies. Further, turbulence could be stratified even on much smaller scales in the vertical, tending to two-dimensional turbulence. Radar observations provided ample evidence for stratified turbulence around and below the tropopause level, with turbulence layers sandwiched between stable layers (e.g. Gage and Green 1978; Roettger and Liu 1978; Gage 1990; Jain *et al.* 2001). These stable layers give rise to enhanced radar signal returns due to Fresnel reflection/scatter (e.g. Gage 1990). In fact, the radar echoes from these layers could be made use of to identify the tropopause level (e.g. Gage and Green 1978).

The value of slope of the turbulence spectrum greater than its isotropic value of $-5/3$ is indicative of anisotropic turbulence (Sidi and Dalaudier 1990). Slope values less than $-5/3$ may be indicative of contamination of the turbulence spectrum by wave disturbances as the spectral slope values of wave disturbances could be less than $-5/3$ (e.g. Sidi and Dalaudier 1990). It may be noted here that, in the present study at any rate, occurrence of slope value $<-5/3$ is rather small compared to the value $>-5/3$.

In the present study, it is found that the slope of the turbulence spectrum is associated with the background horizontal wind speed and the correlation between the two is quite high. Shown in Fig. 3 is the altitude profile of the mean horizontal wind speed for the four seasons. Wind (mean) speed $U$ maximizes in the altitude range 12–15 km in winter seasons and at ~17 km in summer. The maximum value of $U$ is much greater in summer than in winter. The slope $S$ shows similar behaviour to $U$, i.e. it is maximum near the altitude of maximum $U$. However, the maximum value of $S$ is nearly the same in all the four seasons. The altitude region of maximum of $U$ is just below the tropopause level. The cold-point tropopause level, obtained from MST radar observations using the method of Revathy *et al.* (1996) for deriving the temperature profile, is indicated by an arrow marked in the Fig. 3. Jain *et al.* (2001), using the MST radar at Gadanki, reported occurrence of stable layers (stratified turbulence) just below the tropopause level. Thus, it appears that the altitude region of the horizontal wind maximum corresponds to the presence of stratified turbulence (anisotropic turbulence—anisotropy between horizontal and vertical) which could yield a spectral slope value greater than $-5/3$ (Sidi and Dalaudier 1990). Thus, the observation of turbulence spectral slopes in the present study is consistent with the horizontal wind profile and stable layers below the tropopause.

Turbulence is energetically possible when Richardson number, $Ri < 0.25$. As $Ri$ is inversely proportional to the square of wind shear, in Fig. 4, square of wind shear is plotted against $S$ (considering only the mean values) in order to see whether there is any relation between the two. Except in summer 2001 and perhaps in winter 1999, the relationship between these two is not as clear as between $S$ and $U$. It appears that it is the horizontal wind rather than the wind shear which is related to the slope of the turbulence spectrum.

Doppler shifting of gravity-wave spectrum due to background horizontal wind can affect the temporal spectrum of vertical wind. Numerical analysis by Scheffler and Liu (1986) showed that Doppler shifts due to the background horizontal wind tend to redistribute the power in the gravity-wave spectrum extending the spectrum to frequencies greater than $N$. This could affect the slope of the spectrum at frequencies greater than $N$. However, the Doppler shift effects would become significant at high background horizontal wind ($>\sim$20 m s$^{-1}$) (Scheffler and Liu 1986). A perusal of Fig. 3 shows that the background horizontal wind speeds are quite low ($<\sim$15 m s$^{-1}$) in the three winter seasons under consideration, and hence effects of Doppler shifted gravity waves on the temporal spectra of vertical wind (in the inertial sub-range) would be quite small and may be neglected. On the other hand, background wind in the summer season considered reached a peak value as high as $\sim$30 m s$^{-1}$ (Fig. 3). However, it is seen that the peak value of the $S$ is nearly the same as in the winter seasons when the background wind is low. Further, in the present study, care has been taken to avoid any spectra containing the signature of presence of wave-like distinct spectral peaks (in the inertial sub-range). Thus, it is unlikely that the spectral slopes, even in the summer season considered, are greatly influenced by the Doppler shifted gravity waves so can be taken to represent the turbulence spectrum.

## 4. Estimation of turbulence parameters from the three methods

### (*a*) *Variance method*

In the present study, the variance method has been adapted to estimate the turbulence parameters as described in Satheesan and Krishna Murthy (2002). From the spectrum of vertical wind, obtained as described earlier, the vertical wind variance $\overline{w^2}$ has been estimated by integrating the spectral power between $N$ and the Nyquist frequency. Only spectra with $S$ in the range $-1.67 \pm 60\%$ are included for the estimation of the turbulence parameters. As discussed previously, it may be noted here that spectra with higher values of slope may correspond to anisotropic turbulence. In the estimation of variance $\overline{w^2}$ from the temporal spectrum, caution has to be exercised to avoid cases of any strong wave perturbation even at frequencies greater than $N$. For example, Cho *et al.* (1996) reported a strong 2 min period oscillation in vertical wind velocity in the stratosphere (at a frequency greater than $N$) which is attributed possibly to manifestation of Kelvin–Helmholtz instability. Such occurrences are certainly event based and may not be of general occurrence. However, in the present study, care has been taken to avoid spectra contaminated by any such occurrences as mentioned earlier. It may be noted that some of the spectra may be contaminated by the presence of waves (not strong) which are probably revealed in spectral slope values $<-5/3$ (which are of infrequent occurrence).

The turbulent kinetic energy dissipation rate $\varepsilon$ is estimated from $\overline{w^2}$ using the following equation (Hocking 1989):

$$\varepsilon = 6.1F\overline{w^2}\frac{N}{2\pi} = 0.97\overline{w^2}N, \tag{2}$$

with $F = 1$.

It may be noted here that the expression for $\varepsilon$ is obtained assuming isotropic turbulence, i.e. spectral slope $\sim -5/3$. However, even when the spectral slope value is different from (less than) $-5/3$, the expression for $\varepsilon$ corresponding to isotropic turbulence may be assumed to be approximately valid. The factor $F$ (Eq. (2)), which represents the fraction of measured $\overline{w^2}$ that resides in the inertial sub-range (Hocking 1989), can be taken to include any change due to anisotropy of turbulence. The factor $F$ is reported to have a significant range of values (Hocking 1985). In the absence of any definitive information, $F$ is taken as 1 in the present study (Satheesan and Krishna Murthy 2002).

### (*b*) *Power method*

For the power and Doppler spectral-width methods, the oblique antenna beam data, taken from 10 min preceding to 10 min following the 2 h duration of vertical beam data (giving a total of 20 min), are used on each of the observation nights. From the oblique beam data, first the refractive index structure constants $C_n^2$ are obtained for each of the 32 oblique antenna beam data (8 each for north, south, east and west beams) using the expression given by Van Zandt *et al.* (1978) and the relevant radar parameters (Satheesan and Krishna Murthy 2002). These profiles of $C_n^2$ are averaged and the average profiles are used to estimate $\varepsilon$ (Hocking 1985). In the estimation of $\varepsilon$ from $C_n^2$ for the required data of atmospheric pressure and water-vapour pressure profiles, the radiosonde data at a nearby location, Chennai (13.04°N, 80.16°E), are used. For temperature, the data derived from $N$ obtained from MST radar observations using the method of Revathy *et al.* (1996) are used.

### (*c*) *Doppler spectral-width method*

For the estimation of $\varepsilon$ from the spectral-width method, the Doppler spectral widths are obtained for each of the 32 oblique antenna beam data. The corrections due to the beam broadening and shear broadening are applied using equations given in Fukao *et al.* (1994) and the corrected widths are averaged for each night. From this average corrected spectral width, $\varepsilon$ is estimated following Hocking (1989). In the present study, oblique antenna beam data are used instead of vertical antenna beam data for the power and spectral-width methods. This is because the vertical beam data of power and spectral width would be affected by Fresnel scatter and Fresnel reflection as well, in addition to turbulent scatter, whereas the oblique beam data (of power and spectral width) would be largely free from Fresnel scatter and Fresnel reflection effects (Satheesan and Krishna Murthy 2002). On the other hand, for the variance method only the vertical wind (from the vertical beam) variance in the inertial sub-range is considered, thus avoiding contamination by Fresnel scatter and Fresnel reflection. Here it may be noted that in obtaining $\varepsilon$ from the width and power methods using oblique beam data, data of all the four oblique beams are averaged. This implies in effect neglecting any horizontal anisotropy (Nastrom and Tsuda 2001). This aspect is dealt with later.

It may be noted here that the oblique beam data are available for 10 min preceding and following the 2 h data duration of the vertical beam used for the variance method. It is found that the power and Doppler spectral widths for the two 10 min periods are nearly the same. Thus the average values of the two 10 min data periods can be considered to represent $\sim$2 hour average.

## 5. COMPARISON OF $\varepsilon$ FROM THE THREE METHODS

Altitude profiles of $\varepsilon$ by the three methods have been obtained using the data in the four seasons as described and are compared. Typical examples of the $\varepsilon$ profiles obtained by the three methods are shown in Figs. 5 and 6 to illustrate the results of the comparison.

Shown in the Figs. 5 and 6 are the $\varepsilon$ estimates from the variance method using the spectral slopes in the range $-5/3 \pm 60\%$ and $-5/3 \pm 30\%$. It is seen that both these estimates follow the same trend of variation (with altitude). Thus inclusion of a greater range (60%) of values of $S$ does not appear to affect the results in any significant manner.

In the vertical profiles of $\varepsilon$ from the variance method, there are gaps because the spectral slopes are outside the range of slopes ($-5/3 \pm 60\%$) used in the estimation. The gaps generally correspond to regions of high horizontal wind speed (e.g. 27 January 1999 and 12 February 2002) and these are the regions where the spectral slope values are generally high as observed earlier. From the signal-power profiles (shown in the Figs. 5 and 6), it is seen that in general these are also regions of large fluctuations, with signal enhancements indicating stable layers. As pointed out earlier, these indicate the presence of stratified layers (anisotropic turbulence).

A perusal of Figs. 5 and 6 reveals that in the troposphere in the region of low background horizontal wind (i.e. below $\sim$9 km) there is, in general, a good agreement between $\varepsilon$ from the variance and the width methods, whereas that from the power method is different. On the other hand, in the lower stratosphere (18–21 km) under low background horizontal wind conditions, $\varepsilon$ from all three methods differ from each other, with that from the variance method being highest and that from the power method being lowest, except on 12 February 2002 when $\varepsilon$ from the width method is the lowest. In the upper troposphere ($\sim$9–17 km) where the background horizontal wind reaches a maximum with high values, $\varepsilon$ from the variance method is significantly greater than that from the other two methods. However, the magnitude of the difference itself in $\varepsilon$ between the variance and the other two methods does not seem to have a direct relationship with the value of the wind speed (e.g. $\varepsilon$ profiles on 12 February and 3 March 2002 in Fig. 6). In this altitude region, $\varepsilon$ values from the power and the width methods are relatively closer, especially on 28 January 1999.

It may be noted here that there are gaps in $\varepsilon$ values from the width method at some altitudes (for example on 12 February 2002), especially where the background wind speed is high. These gaps occurred because the corrected spectral widths became negative. It appears that the width method does not yield reliable values of $\varepsilon$ when the background wind speed is high, probably because of excessive correction for the beam-width effect. Hocking (1985) pointed out that at times the beam-width correction may render the spectral width attributable to turbulence negative and such cases may occur under weak turbulence. Nastrom and Tsuda (2001) observed significant horizontal anisotropy in the mean spectral widths from radar antenna beams in the meridional and zonal planes. VanZandt *et al.* (2002) noted that horizontal anisotropy may occur at high wind-speed conditions. Nastrom and Tsuda (2001) pointed out the possibility of another correction term, yet unknown, to be applied for spectral width.

It is clear from the above that in the troposphere, where the wind (horizontal) speed is high, $\varepsilon$ values from the three methods differ very significantly. As pointed out earlier, in the altitude region 9–17 km where high (horizontal) winds prevailed, turbulence could be anisotropic as indicated by the spectral slopes of vertical wind. In both the power method and the width method, oblique beam data are used. Thus, turbulence in the horizontal and vertical directions contributes to $\varepsilon$ from these two methods, whereas only turbulence in the vertical direction contributes to $\varepsilon$ from the variance method. The

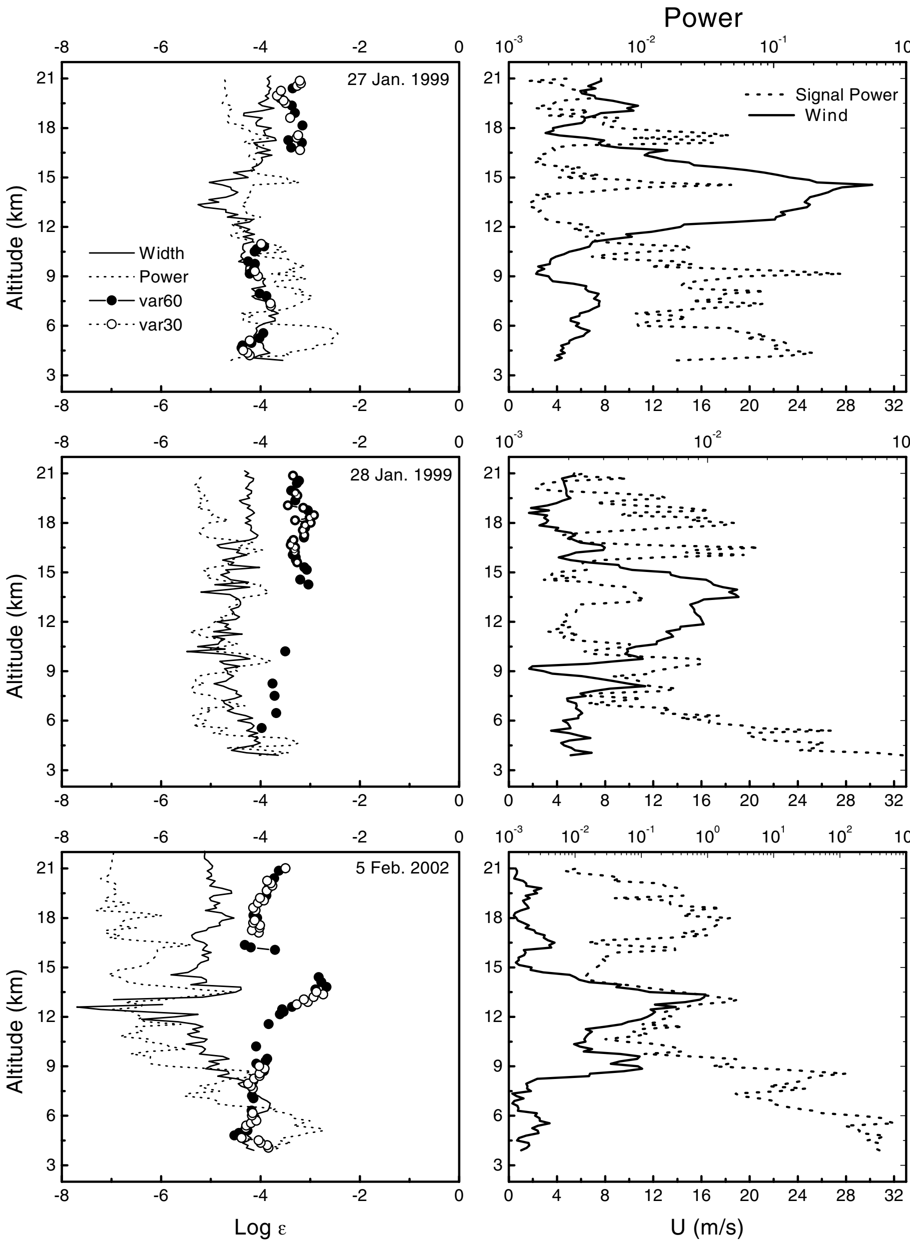


Figure 5. The $\varepsilon$ values (see text) obtained using variance method (open and closed circles, left column) are compared with the $\varepsilon$ obtained by width method (solid line) and power method (dotted line) using the oblique beams on 27–28 January 1999 and 5 February 2002. The right column shows the wind vector on the corresponding days and the radar return signal power.

difference between $\varepsilon$ from the variance method and that from the other two methods in the high background horizontal wind region in the troposphere indicates anisotropy in turbulence (between horizontal and vertical), as is also indicated by the spectral slopes of vertical wind temporal spectra. In the lower stratosphere (18–21 km), it is observed that $\varepsilon$ values from the three methods are significantly different even though the background horizontal wind is low. The stratosphere is a region of high static stability with a positive

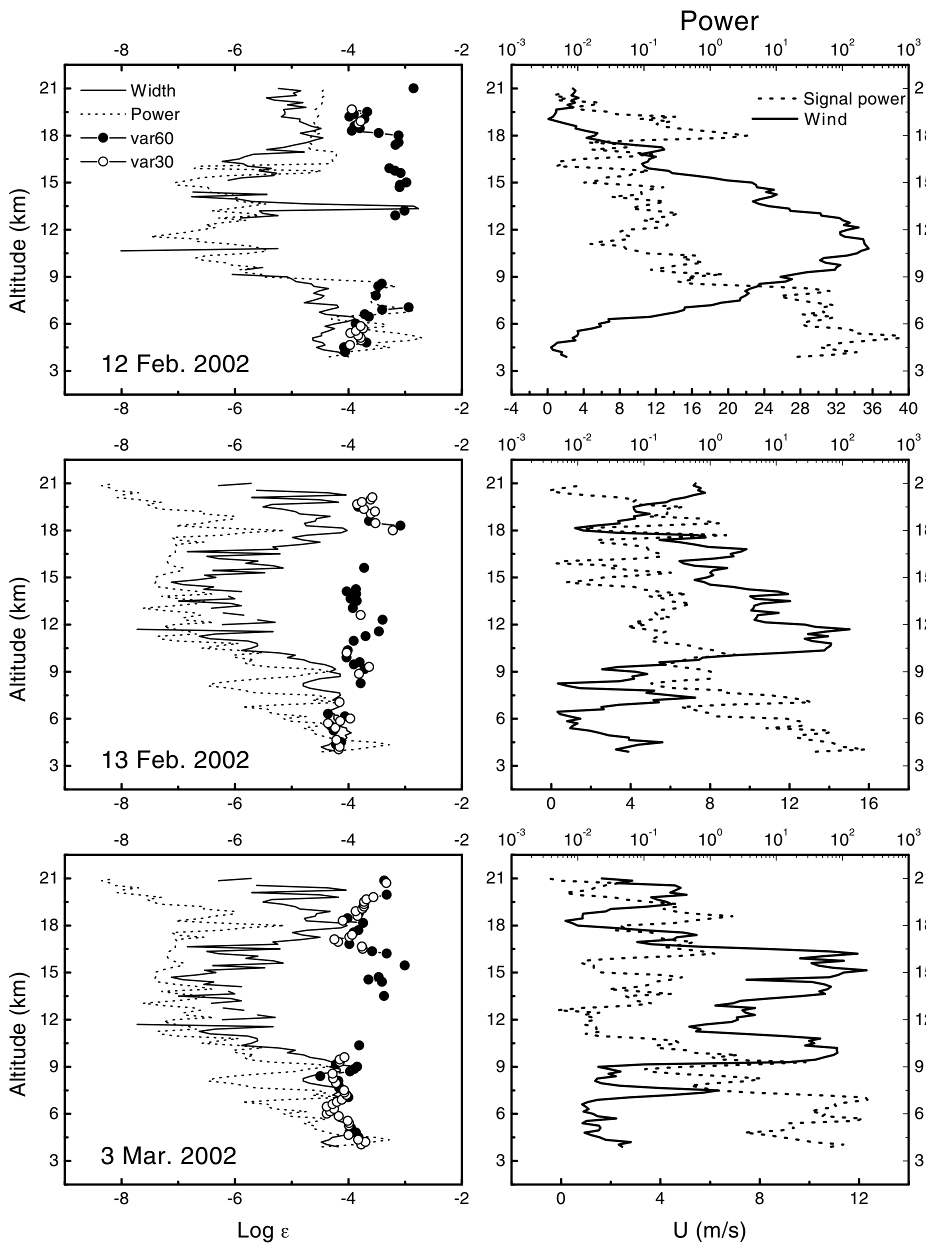


Figure 6. Same as Fig. 5 but for 12–13 February and 3 March 2002.

vertical gradient of temperature, and the lower stratosphere is found to exhibit layers of stratified turbulence (e.g. Jain *et al.* 2001). This is also evidenced by the observed power fluctuations in this region (Figs. 5 and 6). This stratified turbulence, indicating anisotropic turbulence (anisotropy between horizontal and vertical), could be the reason for the observed differences in $\varepsilon$ from the three methods. It is interesting to note that, in the troposphere in the regions of high background horizontal wind, the spectral slopes (of the temporal spectrum of vertical wind in the inertial sub-range) are high, whereas

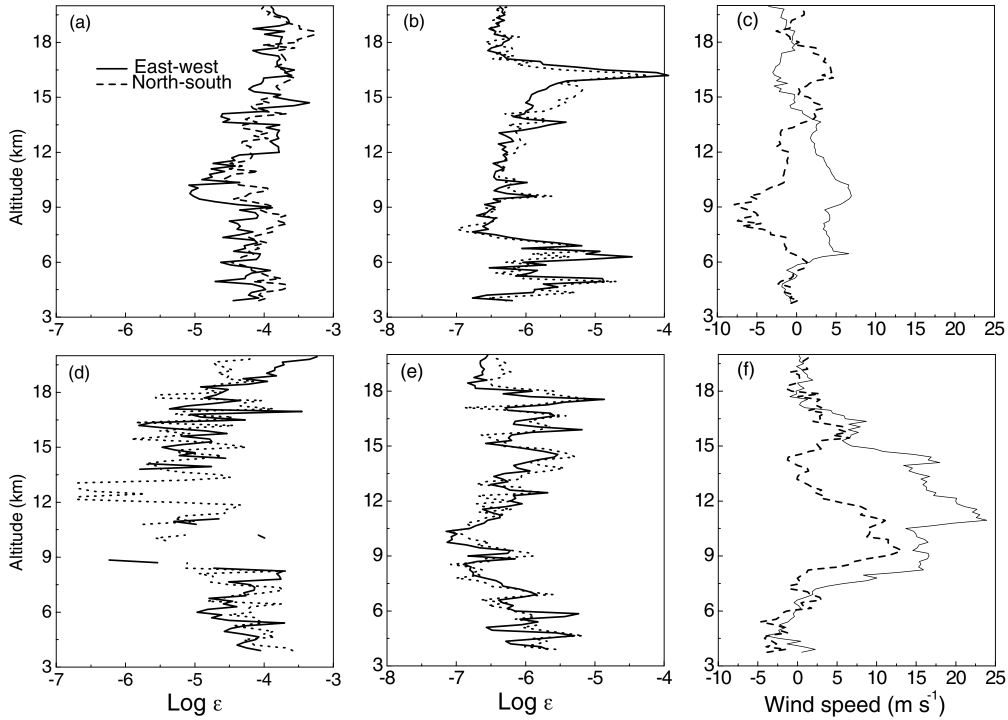


Figure 7. Altitude profiles for (a)–(c) 19 January 1999 and (d)–(f) 11 February 1999: $\varepsilon$ (see text) for east–west and north–south beams, estimated by (a) and (d) the width method and (b) and (e) power methods; (c) and (f) meridional (dashed line) and zonal (continuous line) winds.

in the lower stratosphere they are not. Thus, it appears that anisotropy in regions of high background horizontal wind (in the troposphere) affects the spectral slopes, whereas that in the regions of static stability in the lower stratosphere may not affect the spectral slopes.

If the turbulence is not isotropic, $\varepsilon$ can only be estimated using a model of the anisotropy (Van Zandt *et al.* 2002). In the absence of this, the same expressions as for isotropic turbulence have been used in the present study as for estimation of $\varepsilon$ under anisotropic turbulence. The anisotropic effect can probably be absorbed in the factor $F$ (Eq. (2)).

In the present study, in the estimation of $\varepsilon$ from the width and power methods, average values of width and power for the north–south and east–west beams have been used. Van Zandt *et al.* (2002), using the spectral-width method from dual beam-width radar observations, pointed out anisotropy (horizontal) in turbulence in the zonal and meridional directions when the wind speeds exceed 30 m s$^{-1}$. In view of this, in order to see the effect, if any, of horizontal anisotropy on the estimates of $\varepsilon$ by the width and power methods in the present study, a few cases of estimates corresponding to oblique beam data in the north–south and east–west planes using the spectral-width method under high and low wind conditions are examined: Fig. 7 shows an example of these. Figure 7 also shows the corresponding zonal and meridional wind profiles obtained from the radar observations. The zonal and meridional winds are quite low on 19 January 1999 while they are high on 11 February 1999 reaching a maximum of $\sim$24 m s$^{-1}$ in the mid-troposphere. The values of $\varepsilon$ from the power method for the beam in the north–south and east–west planes are in good agreement on 19 January 1999, when the horizontal wind is low, as well as on 11 February 1999 when the horizontal wind is high. The agreement between $\varepsilon$ values in the north–south and east–west planes from the width method is not as good on 11 February 1999, especially above 18 km. In the altitude region where the zonal wind is high (~8–15 km) $\varepsilon$ could not be estimated by the width method with the east–west beam data as the beam-width correction rendered the spectral width negative. From the above it appears that horizontal anisotropy is not very pronounced even when the horizontal wind is high (~24 m $s^{-1}$). However, it is quite likely that horizontal anisotropy may be exhibited at still higher horizontal winds (Van Zandt *et al.* 2002). Horizontal anisotropy has not been considered in the estimation of $\varepsilon$ from the width and power methods presented earlier, wherein the average values of spectral width and signal power for north–south and east–west beams have been used. However, in view of the above, this may not affect the results of comparison in any significant manner, as the horizontal anisotropy appears to be not very prominent.

It should be pointed out here that the above discussion pertains only to horizontal anisotropy of turbulence. As mentioned out earlier, anisotropy between horizontal and vertical turbulence could be a factor when the horizontal wind is high. The observed substantial differences in $\varepsilon$ from the variance method and the width and power methods, which are mainly in the high horizontal wind regions, could be mainly due to the effect of anisotropy between horizontal and vertical turbulence.

## 6. Conclusions

The values of the turbulence parameter $\varepsilon$ in the troposphere and lower stratosphere estimated using the three methods, namely, the Doppler spectral-width method, the backscatter signal-power method and the wind-variance method are compared. The main conclusions of this study are the following:

(i) The slope of the temporal spectrum of vertical wind in the inertial sub-range attains a maximum in the troposphere in the altitude ranges 12–15 km in winter and 15–18 km in summer, in the region of high background horizontal wind. An exponential relation exists between the slope of the temporal spectrum of vertical wind in the inertial sub-range and the background horizontal wind.

(ii) The $\varepsilon$ estimated by the variance method, in general, matches fairly well with that estimated by the spectral-width method in the troposphere under low background horizontal wind conditions.

(iii) In the troposphere, under high background horizontal wind conditions, $\varepsilon$ from the variance method is generally greater than that from the other two methods. In the lower stratosphere, even under low background horizontal wind conditions, the values of $\varepsilon$ from the three methods are all different. These departures are attributed to anisotropic turbulence under strong wind conditions in the troposphere and under static stability conditions in the lower stratosphere.

## Acknowledgements

The authors would like to express their thanks to the referees for their critical comments and helpful suggestions. The National MST Radar Facility (NMRF) is an autonomous Facility under the Department of Space with partial support from the Council of Scientific and Industrial Research (CSIR). The authors are thankful to the technical and scientific staff of NMRF for their dedicated efforts in conducting the experiments. B. V. Krishna Murthy, would like to acknowledge the CSIR for a grant from Emeritus Scientist Scheme.

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