

SURFACE LAYER TURBULENCE SPECTRA AND DISSIPATION RATES DURING LOW WINDS IN TROPICS

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Abstract. Spectral characteristics of surface layer turbulence in an urban atmosphere are investigated. The observations used for this purpose represent low wind conditions in the tropics. The normalized power spectral shapes exhibit the usual characteristics in the inertial subrange and obey Monin–Obukhov scaling. However, the low-frequency behaviours do not conform to the previous observed relations. For horizontal components, large energy is contained in the low frequencies in contrast to the vertical component where roll-off to zero frequency is faster.

The turbulent kinetic energy dissipation rate estimated from the spectra using Kolmogorov's inertial subrange law is found to be isotropic unlike the velocity variances. The expressions for the dimensionless dissipation rate do not seem to work well in low winds in an urban atmosphere. For the data considered, the dissipation rate exhibits a power law relationship with the mean windspeed and the friction velocity.

1. Introduction

Spectral analysis of the wind and temperature fluctuations are useful in determining spatial and temporal scales. The concept of spectral analysis for atmospheric motions has been used in many studies (Busch and Panofsky, 1968; Kukharets and Tsvang, 1969; Sitaraman, 1970; Kaimal *et al.*, 1972; Caughey, 1977; Moraes, 1988; Roth *et al.*, 1989) for different atmospheric stabilities, varying terrains (roughness) and at various levels in the atmospheric boundary layer. The results of the spectral analysis could also be very useful in parameterizing eddy diffusivities for applications in air pollution problems. For example, the spectral maximum frequency is useful in obtaining diffusion estimates as it can help specify the eddy diffusivity coefficients (Weber *et al.*, 1982). Formulations using this approach have been suggested by Pasquill (1974) and Hanna (1968, 1978) among others.

Another way of parameterizing diffusivities is in terms of turbulent dissipation rate (ϵ) estimated from the inertial subrange of the spectra (Berkowicz and Prahm, 1979; Pasquill and Smith, 1983). The dissipation rate is also used in the calculation of plume rise (Hanna *et al.*, 1982).

Various investigators have shown that over the inertial subrange of frequencies the wind and temperature spectra follow surface layer similarity theory. Kaimal *et al.* (1972) used similarity theory to describe the spectra and cospectra in the surface layer over a broad range of stability conditions. The normalised spectral distribution was observed to depend on z/L (z is the measurement height above

the surface and L is the Obukhov length) with the exception of horizontal spectra in unstable conditions ($z/L < 0$) in the low frequency portion. The vertical velocity (w) spectra, in contrast, showed a definite order in terms of the spectral peak and the low frequency roll-off with z/L . That is, in general, the w -spectra on the low frequency side seem to vary as a function of z/L in the surface layer.

Hogstrom *et al.* (1982) have analysed the similarity characteristics of turbulence statistics of an urban area and its outskirts using local scaling parameters. The turbulence spectra of an unstable suburban atmosphere have been analysed by Roth *et al.* (1989). Several studies have been conducted on turbulence structure in complex terrain with varying complexities (Panofsky *et al.*, 1982; Wang, 1992). However, most of them did not consider low wind conditions, except a few which dealt with very specific terrain such as mountain-valley (Wang, 1992), and rolling hills (Panofsky *et al.*, 1992).

Low wind conditions occur almost everywhere in the world but are most frequent in tropical regions. The turbulence structure of the surface layer in these conditions is poorly understood. The lack of data for these conditions, especially in tropical regions, has been a severe constraint on the study of turbulence structure and spectral analysis. For practical applications in air pollution modeling studies, the results obtained for moderate to strong winds in mid-latitudes still find wide usage.

A portion of the data collected during the field experiments conducted in the campus of the Indian Institute of Technology (IIT), Delhi, India, in 1991–92 has been used to study the spectral characteristics and the rate of turbulent dissipation in the surface layer. For this data set, turbulence statistics were computed in an earlier study (Agarwal *et al.*, 1995). Most of the data utilized in this study represent a typical winter day–night period with weak winds, especially in the late evening and nighttime. This has enabled us to analyse the diurnal variation of certain important parameters such as eddy dissipation rate, turbulent kinetic energy, and friction velocity.

2. Description of Data

The observation site, although flat, has some bushes and small trees in the surroundings. But it lies in the city and represents urban terrain. This site at the campus of the Indian Institute of Technology (IIT) is somewhat inhomogeneous and aerodynamically rough. The micrometeorological tower was located in an open field with sparsely located small trees (< 3 m) to the west of the tower at a distance of about 50 m. The field was open in other directions to a distance of at least 500 m. The main building of the campus was located at a distance of about 600 m north east of the tower. Turbulence measurements were made well above the trees even for the westerly winds. However, the winds are mostly easterly for this period of the year. The roughness length for this site was estimated to be 78 cm in a study conducted in 1987 (Raman *et al.*, 1990) for pre-monsoon conditions. The instrumentation

for the 30 m multi level micrometeorological tower used for the meteorological observations included a sonic anemometer at the 4 m level and slow response sensors such as wind vane, cup anemometers and thermistors at 1, 2, 4, 8, 15 and 30 m levels. In this study, wind (u, v, w) and temperature (T) fluctuations were obtained using the sonic anemometer whereas the gradients of mean windspeeds and temperatures were determined using observations from slow sensors at 2 and 15 m levels.

The data set comprised of 18 hourly test runs with 9 each for stable and unstable conditions. Out of these, 16 belong to a continuous day-night period of 14/15 February 1992 between 1200 and 0700 LST. The sampling frequency for the output of sonic anemometer was averaged to 1 Hz, giving 3600 observations in an hourly test run. In this study, an averaging time of one hour was chosen so that low-frequency energy cut-off is negligible. The mean parameters and the atmospheric stability for different test runs are given in Table I. The data represent essentially low-wind conditions during winter months. Hourly mean windspeed ranged from 0.6 to 3.5 m s⁻¹ at a height of 4 m. A single day's data have been considered for analysis with a view to study, in addition, the diurnal variation of turbulent parameters and their relationship to atmospheric stability and mean windspeed.

3. Results and Discussion

Before analysis, the time traces of each component for all the test runs were examined to detect any spikes, kinks or missing portions in the data. Such problems, although very few, were corrected suitably. For example, the spikes (found at one or two points out of 3600 in very few cases) were replaced by the mean value of the neighbouring points.

Before discussing the spectra, it is good to have an idea of the diurnal variation of the quantities such as mean windspeed (U), temperature (T), friction velocity (u_*) and turbulent kinetic energy (TKE). Figure 1 presents the diurnal variation of T , TKE, U , ϵ and u_* . The average temperature curve shows a maximum around 1500 hrs and decreases thereafter with time. The nighttime part of the curve indicates that surface cooling extends up to 0600 hrs. The radiative cooling seems to dominate at night because of clear sky and weak winds. As a result, turbulence is relatively weak at night. The curves for all the variables show that the conditions at night are characterized by nearly steady-state conditions except around midnight. Evolution of the TKE, U , ϵ and u_* are similar.

3.1. SPECTRA NORMALIZED BY FRICTION VELOCITY

The spectral estimates were obtained using a Fast Fourier Transform technique. Spectra of the three velocity components were computed from the sonic anemometer measurements at 4 m. Observations represented both stable and unstable conditions. Strongly stable and unstable conditions have values of $|z/L|$ beyond the

Table I
Mean parameters and the atmospheric stability for different test runs

Test run	Date	Time (LST)	Mean speed (m s ⁻¹) at 4 m	Temp. (°C) at 4 m	Friction velocity (m s ⁻¹)	z/L
1	13/11/91	1000	0.82	19.72	0.265	-3.29
2	24/01/92	0900	0.61	13.12	0.173	-8.95
3	14/02/92	1200	3.47	14.62	0.615	-0.63
4	14/02/92	1300	3.26	15.60	0.596	-0.69
5	14/02/92	1400	3.18	16.63	0.623	-0.49
6	14/02/92	1500	2.96	16.62	0.590	-0.22
7	14/02/92	1600	2.77	16.62	0.518	-0.09
8	14/02/92	1700	2.12	16.05	0.447	-0.31
9	14/02/92	1800	1.27	14.94	0.317	1.83
10	14/02/92	1900	1.12	13.99	0.262	1.95
11	14/02/92	2000	1.18	13.52	0.244	2.67
12	15/02/92	0000	2.03	11.82	0.386	1.02
13	15/02/92	0100	1.07	10.61	0.286	1.26
14	15/02/92	0300	1.17	9.14	0.278	1.93
15	15/02/92	0400	1.28	8.37	0.267	2.31
16	15/02/92	0500	1.35	7.83	0.325	1.39
17	15/02/92	0600	1.22	7.59	0.322	1.76
18	15/02/92	0700	1.54	7.35	0.347	-0.71

range for which the theoretical basis for Monin–Obukhov scaling is valid. The spectra were normalized by $u_*^2 \phi_\epsilon^{2/3}$ (ϕ_ϵ is dimensionless dissipation rate), as was done by other investigators (Kaimal *et al.*, 1972; Kaimal, 1978). The dimensionless dissipation rate $\phi_\epsilon = \epsilon k z / u_*^3$ (k is von Karman constant) can be expressed as a function of the stability parameter z/L . Following this normalization when spectra are plotted against the non-dimensional frequency $f = n z / U$ (n being the cyclic frequency, and z the observation height), their stability dependence is removed on the high frequency side and they collapse into a single curve and obey $0.3 f^{-2/3}$ for the longitudinal component and $0.4 f^{-2/3}$ (isotropy) for the lateral and vertical components. The mathematical expressions consistent with this Monin–Obukhov scaling are

$$\frac{n S_u(n)}{u_*^2 \phi_\epsilon^{2/3}} = a_1 k^{-2/3} f^{-2/3} \quad (1)$$

$$\frac{n S_{v,w}(n)}{u_*^2 \phi_\epsilon^{2/3}} = b_1 k^{-2/3} f^{-2/3} \quad (2)$$

where a_1 and b_1 are numerical constants.

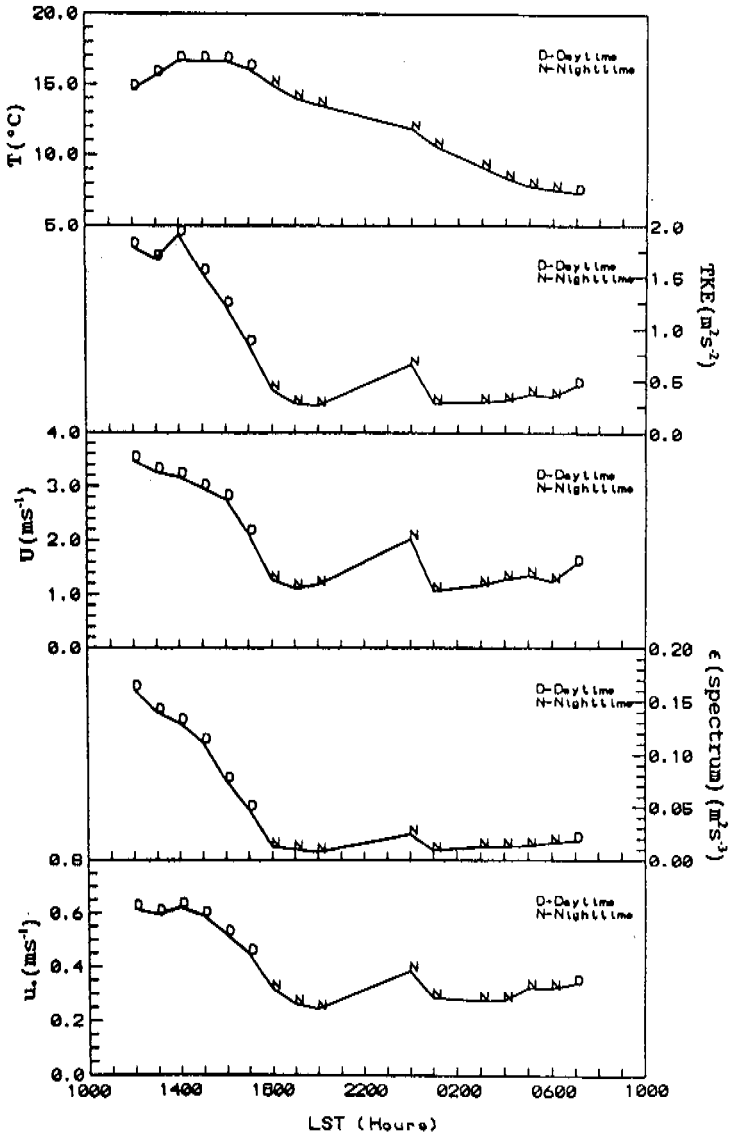


Figure 1. Diurnal variation of hourly mean values of temperature, turbulent kinetic energy, windspeed, turbulent dissipation rate and friction velocity.

Although normalization by u_*^2 is sufficient to make $nS(n)$ non-dimensional, the additional term $\phi_\epsilon^{2/3}$ facilitates the removal of stability dependence in the inertial subrange. On the low frequency side, smoothed spectral curves generally follow a trend with respect to a stability parameter such as z/L (Kaimal *et al.*, 1972). The observations are divided into two groups: nighttime and daytime. The daytime cases are considered to be unstable whereas nighttime cases stable. Figure

2 presents normalized turbulence spectra for longitudinal (u), lateral (v) and vertical (w) velocity components for stable cases. We see that all the spectra collapse on the high frequency end and follow Kolmogorov's $-2/3$ slope (inertial subrange) closely. The low frequency roll-off is faster in v and w spectra which consequently yield a more well defined spectral peak compared to u -spectra. Unlike the results of Kaimal *et al.* (1972), there is no variation with z/L in the low frequency portion for any of the component spectra. Further, present data are for weak wind cases concentrated in a small z/L range. In general, the low frequency ends of the spectra are characteristically noisy (Panofsky *et al.*, 1982). The spectral peaks for various stable runs occur in the range

$$0.02 < f_m^u < 0.08; 0.09 < f_m^v < 0.2; 0.1 < f_m^w < 0.3. \quad (3)$$

The order of increasing spectra energy corresponding to the maximum frequency is w , v , and u components. The spectral energy for the vertical component falls off rapidly below $f \approx 0.1$ and an almost similar trend is observed for the lateral component. The horizontal (especially longitudinal) spectra have most of their energy in relatively low frequencies in contrast to those of vertical component.

The unstable (daytime) spectra for all the three components shown in Figure 3 exhibit the same behaviour as the stable (nighttime) spectra in the inertial subrange. The collapse of all spectra in this range into a single curve is quite good. Again, as in the stable cases, the low frequency behaviour do not seem to depend on z/L . The noise in this portion may be relatively large in unstable spectra. Also, the vertical spectra are again more organized relative to the horizontal ones. The ranges of frequencies for spectral maximum for each of the velocity components are

$$0.02 < f_m^u < 0.05; 0.06 < f_m^v < 0.15; 0.1 < f_m^w < 0.3. \quad (4)$$

These are quite similar to the corresponding ones for stable cases. The order of increasing spectral maximum energy is also the same as in the stable spectra. The spectral amplitudes at low frequencies are larger for the unstable cases (Figure 3) as compared to the stable cases (Figure 2) consistent with the results of Kaimal *et al.* (1972). The spectra of horizontal velocity components display spectral densities that are larger at low frequencies than those of the vertical component. This is a common feature of the horizontal spectra. Our results, like those of Kaimal, indicate that low frequency fluctuations have periods of the order of minutes in unstable air. There is a marked increase in low frequency energy only for the horizontal components. The enhanced amplitudes in low frequencies, at least in stable conditions, may be due to the presence of gravity waves or slow meandering of the airflow. The w -spectra appear to be in local equilibrium at all frequencies.

Dutton *et al.* (1980) have suggested that f_m for w -spectra can be approximated by

$$f_m = 0.183 \quad (z/L < 0.7)$$

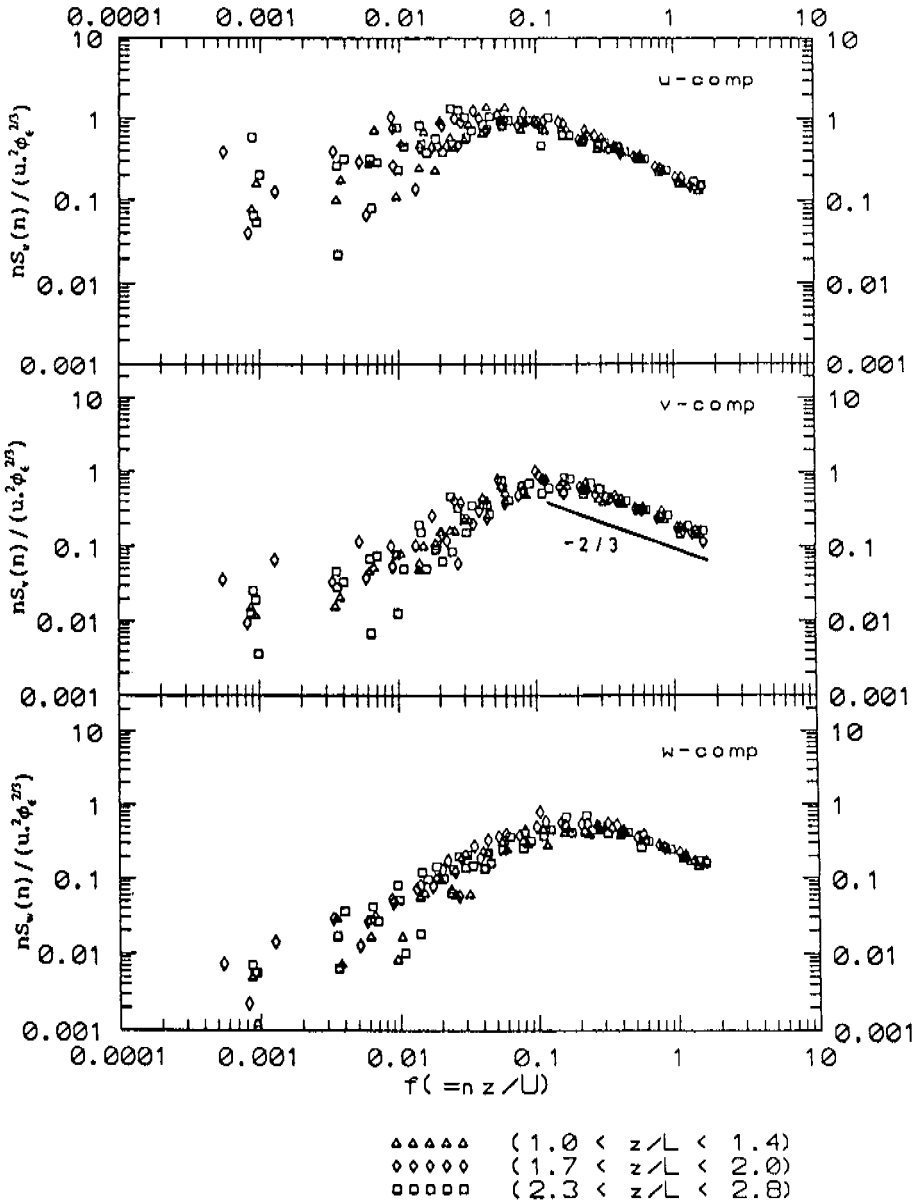


Figure 2. Normalized logarithmic u , v and w spectra plotted against non-dimensional frequency f for stable cases.

$$\begin{aligned}
 f_m &= 0.482 + 0.437z/L \quad (0.7 \leq z/L \leq 0) \\
 f_m &= 0.482 + 0.87z/L. \quad (z/L > 0)
 \end{aligned}
 \tag{5}$$

On comparing our results (Equations (3) and (4)) with these, it is apparent that while the agreement for daytime (unstable) cases is satisfactory, Equation (5)

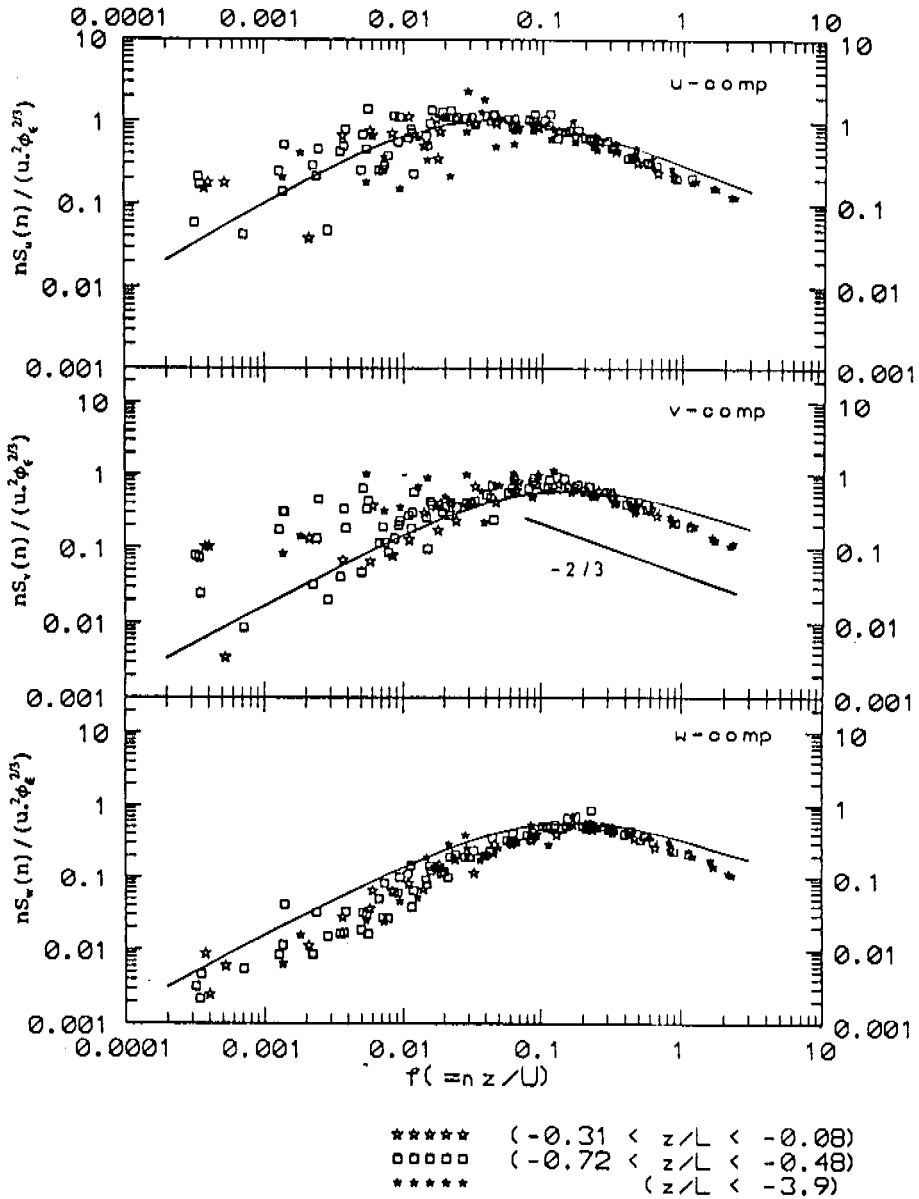


Figure 3. Same as Figure 2, but for unstable cases and the curves shown correspond to empirical relationships (7), (8) and (9) for u , v and w components respectively.

grossly overestimates the nighttime (stable) cases. The reason for overestimation is that low winds are normally characterized by weak wind shear and less TKE. Thus maximum spectral energy is contained in relatively larger scales of motion (i.e., low frequencies).

Thus, the inference is that the spectra of all the three components (stable and unstable) of the velocity in the surface layer obey Monin–Obukhov scaling in the inertial subrange. For low frequencies spectral amplitudes do not seem to depend on z/L . However, for low frequencies our results are in general agreement with the results of Kaimal *et al.* (1972) for both stable and unstable conditions.

Kaimal (1978) and Hojstrup (1982) have proposed models for the spectra of horizontal wind components in unstable conditions which include the parameter z_i (convective boundary-layer height) in addition to z . Kaimal (1978) considered a low frequency (convective) portion which scale with z_i and a high frequency (mechanical) portion which scale with z . Hojstrup's (1982) model describes the spectra by the sum of low and high frequency portions

$$S(n) = S_L(n) + S_H(n) \quad (6)$$

where S_L depends on $f_i = nz_i/U$ and z_i/L , while S_H depends only on $f = nz/U$. Unfortunately, z_i was not measured in our micrometeorological field experiments. Thus, the spectra during convective conditions (Figure 3) could not be compared with the models by Kaimal (1978) and Hojstrup (1982). Nevertheless, the neutral curves have been included to give a rough comparison. The expressions for the horizontal spectra for neutral conditions are (Kaimal *et al.*, 1972)

$$\frac{nS_u(n)}{u_*^2} = \frac{105f}{(1 + 33f)^{5/3}} \quad (7)$$

and

$$\frac{nS_v(n)}{u_*^2} = \frac{17f}{(1 + 9.5f)^{5/3}}. \quad (8)$$

The w -spectra for the convective cases have been compared with the following empirical curve (Kaimal, 1978)

$$\frac{nS_w(n)}{u_*^2 \phi_\epsilon^{2/3}} = \frac{0.4f}{(0.11 + f)^{5/3}}. \quad (9)$$

Most of the spectral amplitudes have slightly low values compared to the empirical relation (Equation (9)). This consistent difference could be due to smaller wind shear associated with the low wind speeds.

The neutral curves (Equations (7) and (8)) for the horizontal components provide good approximations to the observed spectra over a wide range of frequencies. In particular, the spectral peaks are well estimated.

3.2. SPECTRA NORMALIZED BY VARIANCE

Another suitable normalization frequently adopted for atmospheric spectra is by the variance, σ^2 . With a view to obtaining a better insight into this scaling, variation

of all the three velocity components of the spectral density $nS(n)$ normalized by the total variance versus f/f_0 , where f_0 is a frequency related to the characteristic length scale of the parameter, is investigated. The spectra are again obtained for stable (Figure 4) and unstable (Figure 5) conditions separately. The observed spectra at low winds have been compared with Kaimal's (1973) curve

$$\frac{nS_\alpha(n)}{\sigma_\alpha^2} = \frac{0.164f/f_0}{1 + 0.164(f/f_0)^{5/3}} \quad \alpha = u, v, w, T \quad (10)$$

where σ_α^2 is the variance of the parameter α in which f_0 is the value of f where the extrapolated inertial subrange meets the line $nS_\alpha(n)/\sigma_\alpha^2 = 1$. The stable logarithmic spectra are usually expected to follow a roughly symmetric curve which falls off as $f^{-2/3}$ on the high frequency side and as f^{-1} on the low frequency side (consistent with asymptotic spectral behaviour of Equation (10)).

On examining Figures 4 and 5 it is found that Equation (10) provides a reasonably good fit especially in the high frequency range (inertial range) for all the velocity components for stable as well as unstable cases. The prominent features observed are: (1) the spectral peak is slightly underestimated by Equation (10) for the w -component and (2) there is a significant scatter in the low frequency portion of u and v components. For all components, the spectral maximum lies in the range $0.25 < nS_\alpha(n)/\sigma_\alpha^2 < 0.35$ and the corresponding non-dimensional frequency lies in the range $3.0 < f/f_0 < 4.0$ for unstable conditions and $3.5 < f/f_0 < 4.0$ for stable conditions, indicating a broader peak for the unstable relative to the stable cases. Similar results were obtained by Kaimal (1973) for the stable surface layer with a spectral peak of magnitude 0.25 located at $f/f_0 \approx 3.8$.

Based on the information on logarithmic spectral peaks and the intensities of turbulence for the runs considered in this study, it is seen that Lumley's criterion (Kaimal, 1973) for the frozen field assumption is satisfied for stable and convective conditions in low winds. Therefore, it appears that Taylor's hypothesis for frozen turbulence (used for converting time scales to length scales) should be valid for turbulent velocity fluctuations in the surface layer. This result is supported by another study on turbulence statistics by Agarwal *et al.* (1995).

3.3. ESTIMATION OF LENGTH SCALES

Out of several methods for estimating the characteristic length scales, the three most commonly used are based on the power spectrum (Kaimal, 1973). The first method determines an integral length scale, Λ , through the assumption of Taylor's hypothesis. The second length scale λ_m corresponds to the peak of the logarithmic power spectrum. The third scale l is derived from the considerations of available kinetic energy and the energy dissipation rate. All three length scales can be determined from the power spectrum. The length scale Λ is determined by extrapolating the low frequency roll-off to zero frequency, λ_m from the logarithmic spectrum

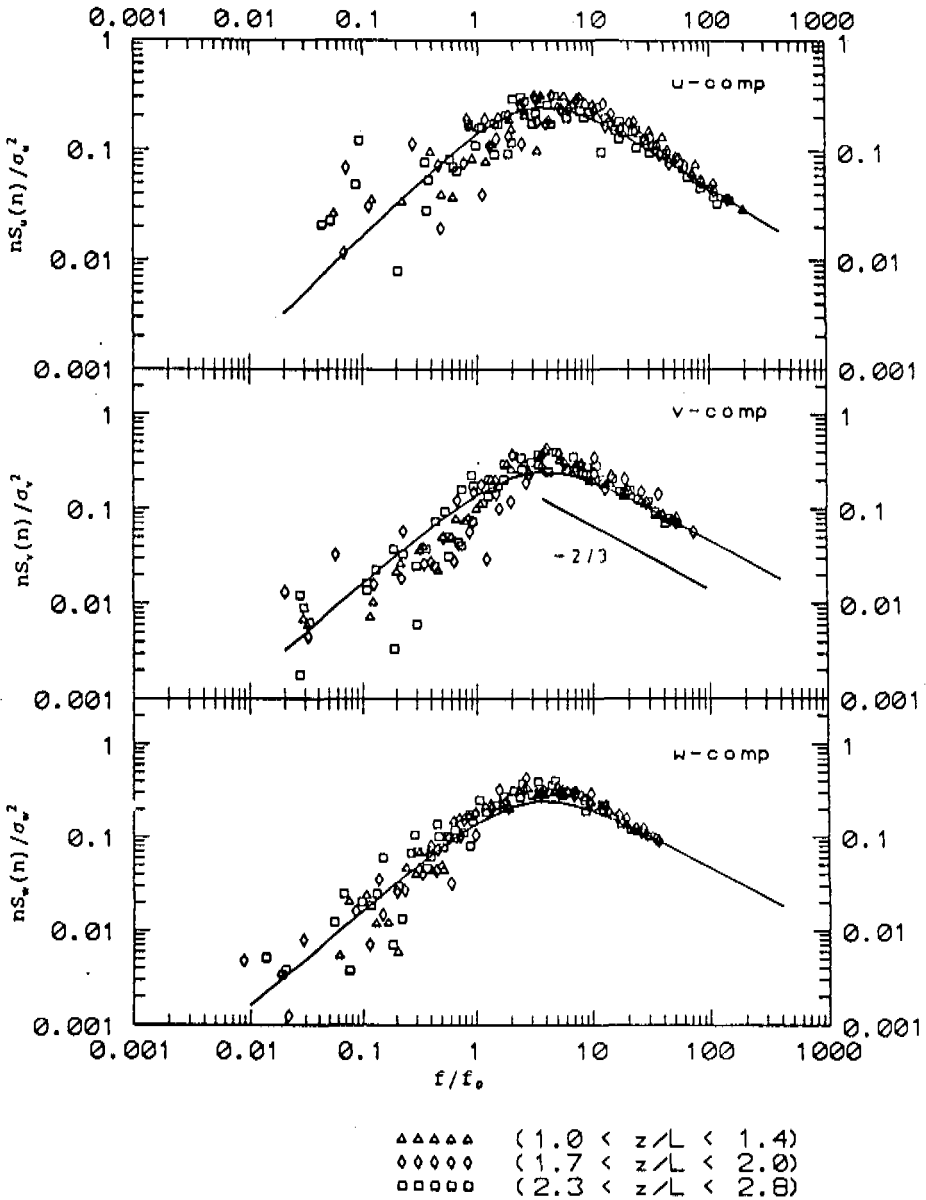


Figure 4. Logarithmic spectra of u , v and w , normalized by their respective variance, plotted against the modified f scale for stable cases. Curves shown correspond to empirical relation (10).

and l from the inertial subrange. With some approximations, all the three length scales can be expressed in terms of z/f_0 (Kaimal, 1973) as below:

$$\Lambda_\alpha = 0.041 \frac{z}{(f_0)_\alpha}; \quad \alpha = u, v, w$$

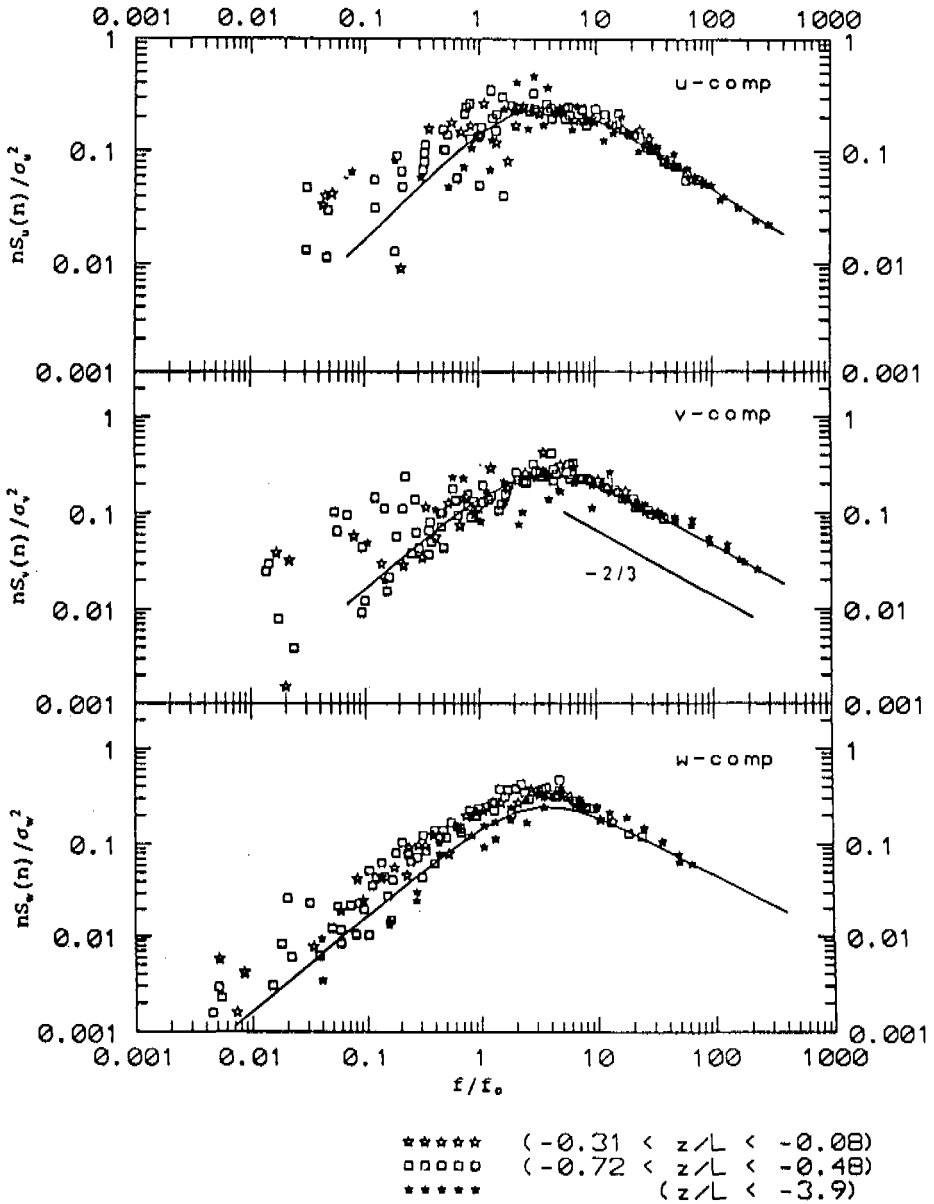


Figure 5. Same as Figure 4, but for unstable cases.

$$(\lambda_m)_\alpha = 0.26 \frac{z}{(f_0)_\alpha} \approx 2\pi\Lambda_\alpha; \quad \alpha = u, v, w$$

$$l_u = 0.056 \frac{z}{(f_0)_u}; \quad l_v = 0.087 \frac{z}{(f_0)_v}; \quad l_w = 0.087 \frac{z}{(f_0)_w}.$$

Table II

Estimated values of f_0 for all the test runs for different wind components

Test run	f_0^u	f_0^v	f_0^w
1	0.0174	0.0092	0.0339
2	0.0099	0.0174	0.0455
3	0.0105	0.0224	0.0592
4	0.0110	0.0252	0.0753
5	0.0073	0.0201	0.0690
6	0.0087	0.0174	0.0720
7	0.0089	0.0242	0.0550
8	0.0100	0.0262	0.0615
9	0.0093	0.0267	0.0639
10	0.0113	0.0366	0.0593
11	0.0077	0.0303	0.0618
12	0.0076	0.0270	0.0629
13	0.0112	0.0223	0.0447
14	0.0218	0.0323	0.0472
15	0.0156	0.0303	0.0470
16	0.0120	0.0260	0.0388
17	0.0174	0.0332	0.0478
18	0.0153	0.0304	0.0477
Mean values	f_0^u	f_0^v	f_0^w
Daytime	0.011 ± 0.003	0.0214 ± 0.006	0.0577 ± 0.013
Nighttime	0.0135 ± 0.005	0.029 ± 0.004	0.0526 ± 0.009

The values of $(f_0)_\alpha$ and the three length scales, are listed for all runs in Tables II and III respectively. Based on the mean values of f_0 it can be concluded that $f_0^w \approx 2f_0^v \approx 5f_0^u$ for unstable cases and $f_0^w \approx 2f_0^v \approx 4f_0^u$ for stable cases. The average values of f_0 for stable cases for u , v and w components based on Kansas data as reported by Kaimal (1973) are

$$0.0354 \pm 0.015, \quad 0.114 \pm 0.05, \quad 0.224 \pm 0.086$$

respectively. These values correspond to measurements at 5.66 m and are 3–4 times higher than our values at 4 m. Also, the dimensionless characteristic length scales shown in Table III are larger than those for Kansas data. Although there are numerous differences between the two sets of observations, the exact reason for the increase in the values of length scales is not clear. However, the general behaviours of the length scales are quite similar to those obtained by Kaimal (1973), i.e., (1) the order of decreasing non-dimensional length scales is u , v , and w , (2) the variation in length scales is maximum in the u component. It decreases for v and w .

Table III
Dimensionless length scales for each test run

Test run	Λ_u/z	Λ_v/z	Λ_w/z	l_u/z	l_v/z	l_w/z
1	2.36	4.47	1.12	3.22	9.49	2.57
2	4.16	2.36	0.90	5.68	4.99	1.91
3	3.9	1.83	0.69	5.32	3.88	1.47
4	3.72	1.63	0.54	5.08	3.46	1.15
5	5.63	2.04	0.59	7.69	4.32	1.26
6	4.72	2.35	0.57	6.45	4.99	1.21
7	4.6	1.69	0.75	6.28	3.60	1.58
8	4.09	1.57	0.67	5.59	3.32	1.41
9	4.4	1.54	0.64	6.01	3.26	1.36
10	3.62	1.12	0.69	4.94	2.37	1.47
11	5.3	1.35	0.66	7.24	2.87	1.41
12	5.37	1.52	0.65	7.34	3.22	1.38
13	3.67	1.84	0.92	5.01	3.91	1.95
14	1.88	1.27	0.87	2.57	2.69	1.84
15	2.63	1.35	0.87	3.59	2.87	1.85
16	3.42	1.58	1.06	4.67	3.35	2.24
17	2.36	1.24	0.86	3.22	2.62	1.82
18	2.68	1.35	0.86	3.66	2.86	1.82

3.4. TURBULENT KINETIC ENERGY AND ITS DISSIPATION RATE

The mean turbulent kinetic energy (TKE) has been computed using velocity fluctuations measured by sonic anemometer:

$$TKE = \frac{1}{2}(\overline{u^2} + \overline{v^2} + \overline{w^2}) \quad (11)$$

where u , v , and w' are the turbulent fluctuations of velocity in the longitudinal (along-wind), lateral (cross-wind) and vertical directions and the overbar represents an average over the hourly test runs. It is assumed that the duration of each of the test runs (hourly) is long enough to include all of the eddy energy.

The ratio of the horizontal to the vertical velocity variances is found to vary between 3 and 4 in the stable cases whereas it is relatively large (3 to 7) in unstable cases. Our ratios lie in between those of Caughey *et al.* (1979) and Louis *et al.* (1982); Caughey *et al.* reported the ratios to be 2 or 3 whereas Louis *et al.* found the ratios to be 10 or more. Caughey *et al.*'s ratios are smaller because they eliminated low frequency peaks in their energy spectra. On the other hand, Louis *et al.* attributed their large ratios to horizontal undulating motions. The anisotropy in TKE exists in our case as well but is not as large as obtained by Louis *et al.* (1982) for stable flows.

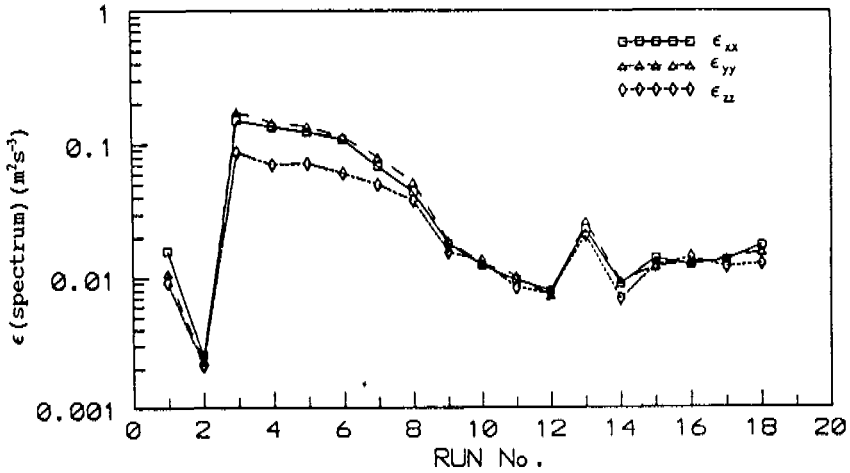


Figure 6. Longitudinal, transverse and vertical dissipation rates for all the test runs considered.

The dissipation rate (ϵ) of TKE for each test run has been obtained from the inertial subrange of the spectra of velocity components using Kolmogorov's law (Taylor's hypothesis of frozen turbulence has been used to convert wave numbers to frequencies),

$$S(n) = \alpha_1 \left(\frac{U}{2\pi} \right)^{2/3} \epsilon^{2/3} n^{-5/3} \quad (12)$$

where α_1 is a universal constant (Kolmogorov constant) estimated from various experiments to be about 0.5. Figure 6 represents the dissipation (ϵ) computed from the spectra of three velocity components for each of the test runs. Anisotropy is apparent in only a few unstable cases, where anisotropy in the velocity variances also exists. Like the results of Louis *et al.* (1982), the dissipation rate is found to be isotropic. Their results are based on the measurements at 18 m and 72 m whereas our measurements were taken at a much lower height 4 m. They observed relatively smaller magnitudes of the TKE and the turbulent dissipation rate. This is quite understandable because the TKE is maximum near the surface and decreases with height and; further, the turbulent dissipation is directly proportional to the TKE.

The estimates of dissipation rate (ϵ) are obtained from the arithmetic mean of longitudinal (ϵ_{xx}) and lateral (ϵ_{yy}) components which represent different estimates of the same quantity. These are computed in order to test the existing relationships (obtained from Kansas data) (Kaimal *et al.*, 1972) of the dimensionless dissipation rate ϕ_ϵ for low wind conditions. Figure 7 shows a comparison of ϵ estimated from the spectra and the one computed using

$$\epsilon = \frac{u_*^3}{kz} \phi_\epsilon \left(\frac{z}{L} \right) \quad (13)$$

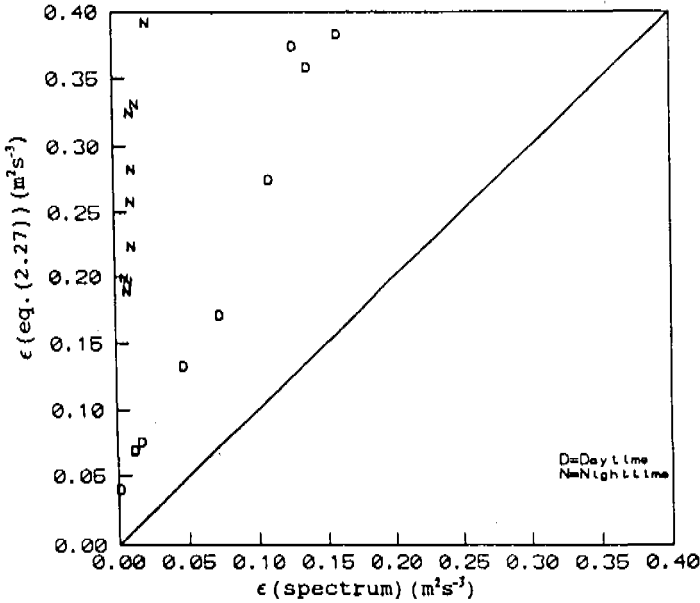


Figure 7. Comparison of dissipation rates estimated from spectra and from expressions for dimensionless dissipation rate based on Kansas data.

where ϕ_ϵ is taken from Kaimal *et al.* (1972). The figure reveals that the trend for unstable cases, although linear, is somewhat different from those based on Kansas data. For stable conditions, the correlation is very poor. Therefore, it appears that the expressions for dimensionless dissipation rate based on Kansas (the site is flat and uniform) data do not work in low wind conditions in an urban (relatively rough and non-homogeneous) atmosphere. Although not explicitly mentioned, it is believed that the Kansas data do not represent low winds.

The turbulent dissipation rate (ϵ) calculated from the power spectra is examined for its variation with the mean wind speed, U and the friction velocity, u_* . The variation of ϵ with U seems (Figure 8) to obey a power law $\epsilon = 0.01U^{2.14}$ in unstable conditions. The exponent is close to 2 and the term U^2 can be identified analogous to TKE which indicates that more TKE implies more dissipation. This seems quite justified. For stable cases, the variation is almost linear although most of the observations in the present study are clustered in a small range of wind speed.

Figure 9 gives the variation of ϵ with u_* . It is seen that all the low wind observations irrespective of stability fall close to a power law curve which can be written as $\epsilon = 0.6u_*^{3.14}$. Referring to Equation (13) in relation to this curve fit, it is apparent that ϕ_ϵ from Kansas data does not fit the present observations in low winds.

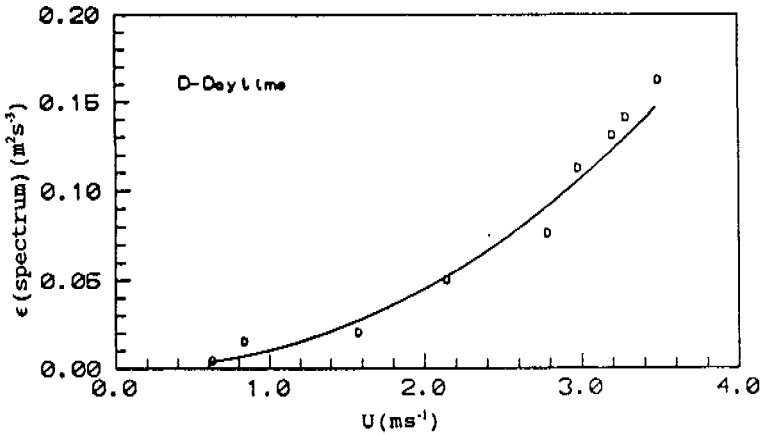


Figure 8. Variation of turbulent kinetic energy dissipation rate with mean windspeed for unstable cases. Curve shown corresponds to the best fit power law.

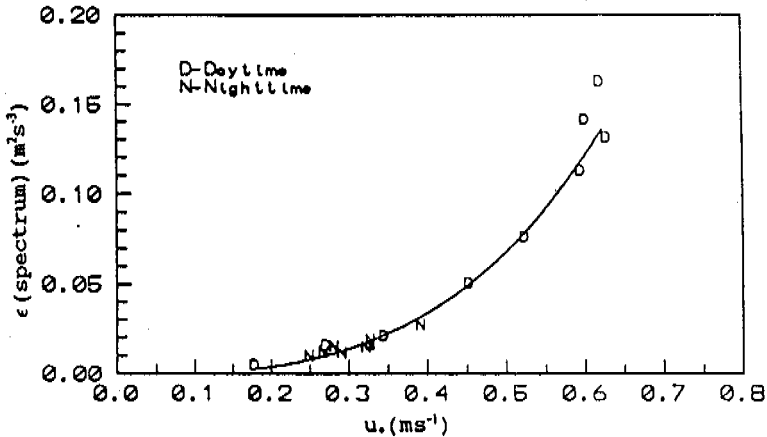


Figure 9. Variation of turbulent kinetic energy dissipation rate with friction velocity. Curve shown here corresponds to the best fit power law.

4. Summary and Conclusions

The normalized spectra of turbulent velocity fluctuations under low wind conditions in convective as well as stably stratified surface layers collapse onto universal curves when plotted against a dimensionless frequency. The inertial subrange of both the horizontal and the vertical spectra follow Monin-Obukhov similarity whereas their energy-containing regions do not present a clear picture especially for the horizontal spectra. Although the results are based on limited data and possess some inherent uncertainties, they show a rough agreement with the results of Kaimal *et al.* (1972). The spectral peaks for the longitudinal component fall within an octave whereas

for the lateral and vertical components the range for the spectral peak is relatively narrow.

The dissipation rate for turbulent kinetic energy estimated from the inertial subrange exhibits isotropy similar to the results of Louis *et al.* (1982). The results indicate, using an indirect approach, that the expressions for dimensionless dissipation rate ϕ_ϵ based on Kansas data are invalid for low wind observations in the tropics. More data need to be generated to obtain meaningful expressions for low wind situations. The dissipation rate is seen to vary as a power law with mean windspeed and friction velocity. The values of the exponents are consistent with theoretical considerations.

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