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SPECTRA OVER COMPLEX TERRAIN

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ABSTRACT

Spectra have been measured over land downwind of a water surface, over hilltops and escarpments, and over rolling farmland. The following hypotheses can be used to explain the differences between these spectra. (1) For wavelengths short compared to the fetch over the new terrain, spectral densities are in equilibrium with the new terrain. (2) For wavelengths long compared to this fetch, spectral densities remain unchanged if the ground is horizontal. If the flow is over a steep hill, the low-frequency structure is modified by distortion of the mean flow, with the longitudinal component losing energy relative to the lateral and vertical components.

Because vertical-velocity spectra contain relatively less low-frequency energy than horizontal-velocity spectra, energetic vertical-velocity fluctuations tend to be in equilibrium with local terrain.

I. INTRODUCTION

Spectra of velocity components in the surface layer over uniform terrain are now generally well understood, thanks largely to the efforts of Kaimal and his co-workers (1972, 1973, 1976, and 1978). Briefly, spectra of the vertical-velocity component obey Monin-Obukhov scaling, with possibly some weak influence of z_i, the height of the lowest inversion, at low frequencies. In contrast, low-frequency portions of horizontal-velocity spectra scale with z_i, with only the highest frequencies obeying Monin-Obukhov and Kolmogorov scaling.

In most applications of velocity spectra to practical problems, however, the ground is not flat and uniform. Wind turbines are usually sited close to the sea or on hills, and towers and bridges are built in various types of complex terrain. In this paper, we discuss spectra of the velocity components obtained over land downwind of a water surface, over hilltops and escarpments, and over rolling farmland.

The following propositions explain features in the wind spectra common to these diverse situations. (1) High-frequency fluctuations (wavelengths much shorter than the fetch over the changed terrain) respond rapidly to changes imposed on the flow by streamline distortion and surface geometry. Thus high-frequency turbulence is always in a state of quasi-equilibrium and the spectral shape conforms to that observed over flat, uniform terrain. (2) Low-frequency fluctuations (wavelengths much longer than the fetch over changed terrain) are affected by upstream terrain characteristics and streamline distortion.

II. SPECTRA OVER LAND DOWNWIND OF WATER

When air moves from water to land, many surface properties change. Here we restrict ourselves to near-neutral conditions, where the most important change is that of roughness length z_0. Over water, z_0 is typically 0.01 cm, and over the land considered here it is 3 cm.

Downwind of a line of roughness change an internal boundary layer develops, as first shown by Elliott (1958). The air within this layer is modified by the new surface, whereas the air above it essentially retains its upstream properties. When the air is neutrally stratified and flowing at right angles to the line of roughness change, the geometry of the interface between these two regions is well known. Elliott found that the height h of the interface is given by

$$\frac{h}{z_0} = a\left(\frac{x}{z_0}\right)^{0.8} , \tag{1}$$

where $a = 0.75 + 0.03 \ln (z_0'/z_0)$, and z_0 and z_0' are upwind and downwind roughness lengths, respectively.

In this section we discuss the influence of such a roughness change on velocity spectra obtained during the Risø 78 experiment conducted at the Risø National Laboratory, Denmark, 1978. We shall compare spectra obtained at two masts. Mast 0 was in water about 100 m from the shoreline, and mast 2a was about 50 m inland, with the over-land fetch varying from 50 to 70 m, depending on wind direction. Mast 0 was instrumented at a height of 2 m and mast 2a at heights of 2, 4, 8, and 12 m with Norman three-dimensional drag anemometers (Perry et al. 1978).

Figure 1 shows u-velocity spectra at a height of 2 m for near-neutral conditions over water (mast 0) and 70 m downwind of the shore-line (mast 2a). The abscissa is the normalized frequency $f = nz/V$, where n is frequency in hertz, and V is the local mean wind speed. The ordinate is the spectral density multiplied by n. The figure shows that there is no significant difference between the two spectra at low frequencies. However, the increased roughness of the land has produced increased spectral densities at high frequencies. These results agree with the two propositions stated in Section 1.

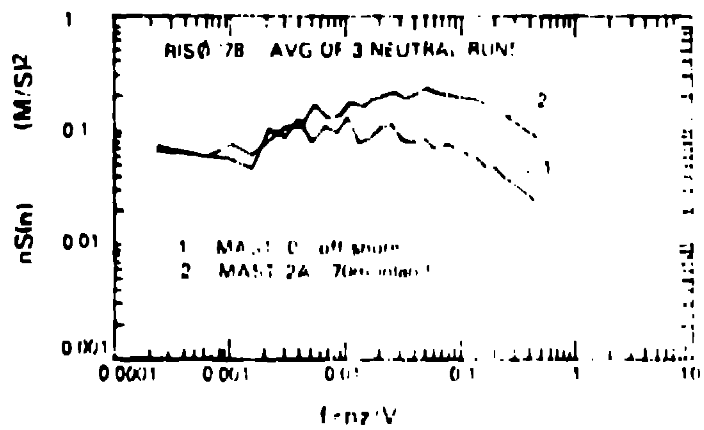

Fig. 1. Average spectra for the u component in near-neutral conditions 2 m above water and land at Risø.

We have had some success in modeling this kind of spectral adjustment by hypothesizing that spectral densities respond to a step-change in surface stress (which accompanies a change in z_0) like a first-order mechanical system, such that

$$\frac{dS(k_1)}{dt} = \frac{S_1(k_1) - S(k_1)}{\tau} . \tag{2}$$

Here $S(k_1)$ is the spectral density at the one-dimensional wave number k_1 ($= 2\pi n/V$ by Taylor's hypothesis), and $S_1(k_1)$ is the spectral density that would be in equilibrium with its environment. The quantity τ is a response time that will be short for small eddies (large k_1) and long for large eddies (small k_1). This simplified approach neglects contributions to $S(k_1)$ from transport terms and the transfer of energy between wave numbers.

If we make the crude assumption that neither V nor $S_1(k_1)$ change significantly downwind of the roughness change, the integration of Eq. (2) is straightforward. We let S_0 equal the spectral density upwind of the terrain change and integrate from $x = 0$ to some arbitrary distance x to obtain

$$\frac{S_1 - S}{S_1 - S_0} = e^{-x/V\tau} \quad , \tag{3}$$

which applies at a particular wave number k_1. Equation (3) has been used to estimate values for τ at different wavelengths from the data in Fig. 1. S_0 was obtained from the over-water spectrum at 2 m; S_1/S_0 was assumed equal to the ratio of the downstream to upstream spectral densities at very high frequencies.

Lumley and Panofsky (1964, p. 85) use dimensional reasoning to define a time scale $\tau = [k_1^3 S(k_1)]^{-1/2}$ for an eddy of size k_1. Using Kolmogorov's law for the inertial subrange, it follows that, for small eddies,

$$\tau_{small} \propto \epsilon^{-1/3} k_1^{-2/3} \quad , \tag{4}$$

where ϵ is the local dissipation rate. For large eddies we assume that the time scale is defined by the wavelength and an eddy velocity, taken to be the standard deviation of the longitudinal component σ_u, such that

$$\tau_{large} \propto (k_1 \sigma_u)^{-1} \quad . \tag{5}$$

Equations (4) and (5) are in general agreement with recent measurements of narrow-band spectral decay by Kellogg and Corrsin (1980).

Figure 2 shows the variation of τ with $(\sigma_u k_1)^{-1}$ at the 2-m height. According to Eqs. (4) and (5), τ should be proportional to k_1^{-1} at low k_1 and proportional to $k_1^{-2/3}$ for large k_1, relationships which, in view of the rather crude nature of the hypothesis and assumptions, are borne out surprisingly well by the data.

Figure 3 shows average spectra of the three velocity components in near-neutral conditions. In each case, the spectra on the inland mast at heights of 6 m and 12 m are not significantly different from upstream spectra at 2 m. This is consistent with the fact that these heights are above the interface (h ≈ 6 m at mast 2a).

At a height of 4 m, just below the interface, a slight increase is apparent in the high-frequency portion of the longitudinal spectrum, but this increase is not easily distinguished in the other two components. At the 2-m height, the high-frequency portion of all components has been strongly increased above their upstream levels.

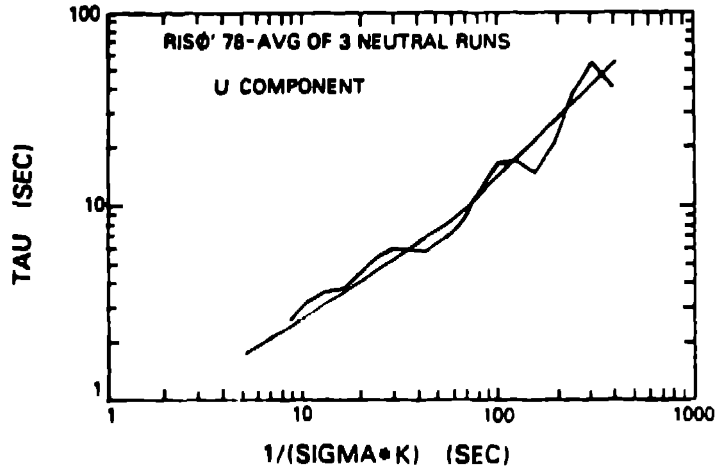

Fig. 2. Response time τ as a function of $1/\sigma_u k_1$. Smooth line is hypothetical.

The low-frequency ends of the spectra are characteristically noisy but clearly are unaffected by the roughness change.

The normalized frequency above which the spectral density is amplified is about 10^{-2} in the case of the longitudinal spectrum but only 10^{-1} for the lateral and vertical components. This difference is particularly important in the vertical velocity component. Because energy in the vertical component falls off rapidly below $f \approx 10^{-1}$, the total variance would be expected to be about the same as that over uniform land of similar roughness.

The difference in the behavior of the three components may be due to the mechanical origin of the increase of turbulence over the rough terrain, so that the longitudinal component is modified first.

III. SPECTRA OVER HILLTOPS AND ESCARPMENTS

The properties of the turbulence on hilltops in neutral conditions are now quite well understood, with theoretical studies (Jackson and Hunt 1975; Mason and Sykes 1979) and observations (Bradley 1980; Mason and Sykes 1979) in good agreement. Generally there is an "inner region" of strong shear in which variances and stress decrease with height and whose thickness depends on the hill length scale and surface roughness. Above this layer is an "outer region" with weak shear where the mean streamlines follow potential flow theory, and changes in the turbulence may be obtained by rapid-distortion theory (Jackson and Hunt 1975).

First we consider spectra obtained in neutral conditions on the top of Black Mountain, Canberra, a hill rising 170 m above the surrounding plain and with a half-length of 275 m. The depth of the inner region at the hilltop was observed to be 28 m, in close agreement with the theoretical value. Site description and experimental details are given by Bradley (1980).

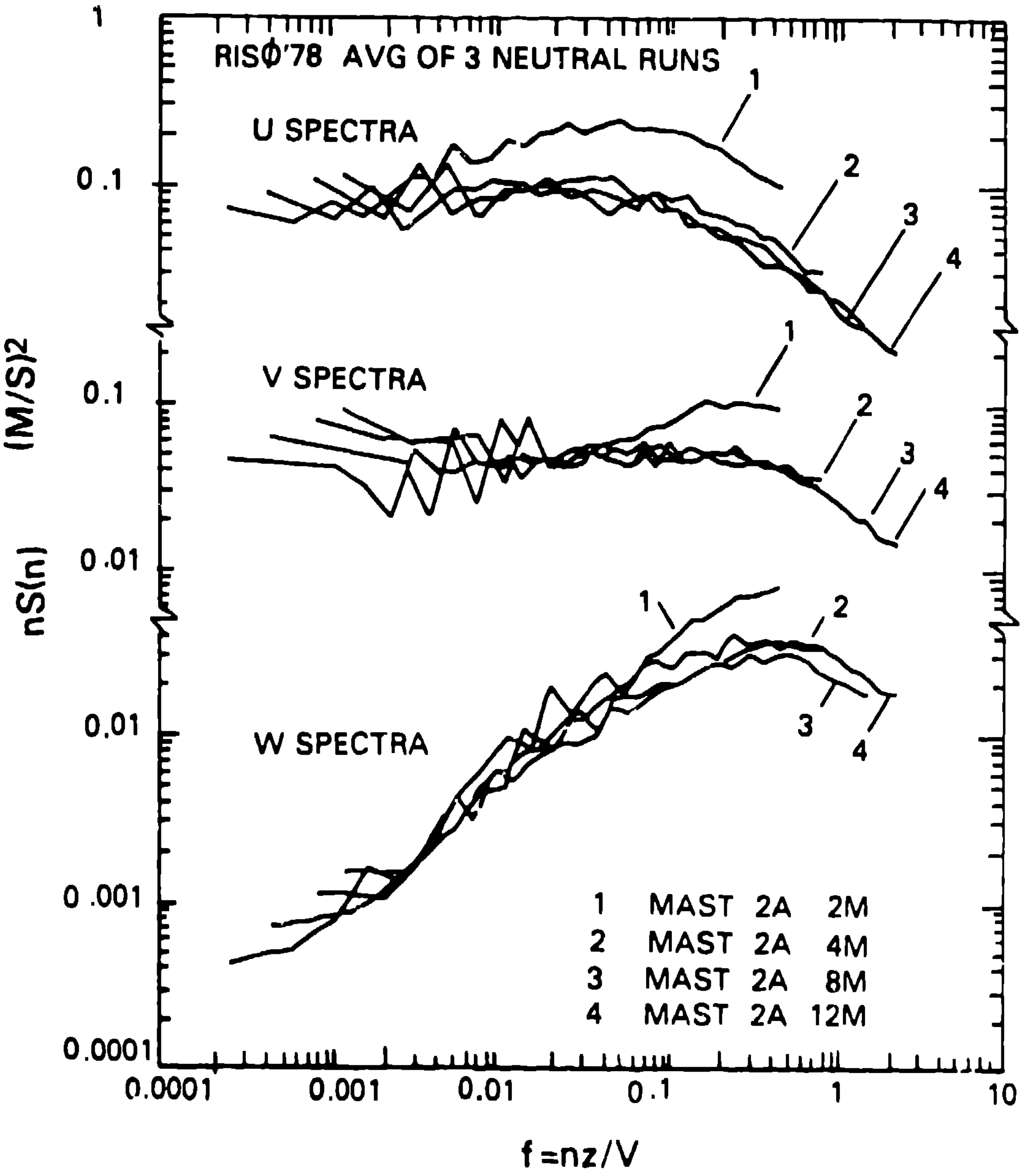

Fig. 3. Average spectra for the three wind components in near-neutral conditions, at heights of 2, 4, 8, and 12 m, 70 m inland at Risø.

Because upwind reference spectra are not available, we used flat-terrain spectral models suggested by Kaimal et al. (1972) to approximate the upstream spectrum; we used models for the neutral limit approached from the unstable side. To compare the model with observed spectra, we must use a scaling law that accounts for the change of stress with height.

We appeal to our first proposition in Section I and to inertial subrange theory to devise a normalization scheme appropriate to hill-tops. Because high-frequency fluctuations adjust rapidly to changes in the flow, we expect spectral densities to be given by Kolmogorov's law

$$S(k_1) \propto \varepsilon^{2/3} k_1^{-5/3} \quad . \tag{6}$$

We now assume that in the inner region, where shear and Reynolds stress are large, local production of turbulence approximately balances the dissipation, or

$$\varepsilon = - \overline{u'w'} \frac{\partial V}{\partial z} \quad , \tag{7}$$

where $\overline{-u'w'}$ is the local Reynolds shear stress and the primes indicate velocity fluctuations. The local shear $\partial V/\partial z$ is more difficult to derive, but at both the Risø and Black Mountain sites we find that the inner region wind profile is closely logarithmic. Furthermore, an "apparent" friction velocity u_* computed in the usual way from this profile agrees remarkably well with the friction velocity obtained by extrapolation downwards of directly measured Reynolds stresses. Thus, in the usual formulation with Von Karman's constant ,

$$\frac{\partial V}{\partial z} = \frac{u_*}{kz} \quad . \tag{8}$$

Combining Eqs. (6), (7), and (8) and converting from wave number to frequency ($k_1 = 2\pi n$) leads to the following nondimensional expression for the energy spectrum function in the inertial subrange

$$\frac{nS(n)}{(\overline{-u'w'}\, u_*)^{2/3}} = af^{-2/3} \quad , \tag{9}$$

where a is 0.3 for the longitudinal component and 0.40 for the lateral and vertical components. Note that over flat terrain, where $\overline{-u'w'}$ is invariant with height near the ground, the left hand side of Eq. (9) reduces to $nS(n)/u_*^2$, the neutral-atmosphere scaling used by Kaimal et al. (1972).

Figure 4a shows spectra of the u component at two heights above the top of Black Mountain; the 9-m level is within the inner region, and the 86-m level is in the outer region. If we disregard the instrumental filtering at very high frequencies, we see that normalized spectral densities at the 9 m height coincide with those of the model for $f > 0.3$. This range of f-values corresponds to wavelengths of 30 m or less, which are short compared to the length of the hill.

In the low-frequency end of the spectrum, normalized spectral densities fall significantly below the model curve, which we have assumed represents the undisturbed upwind spectrum. We expect that this "deficit" is due to memory of a smaller u_* upstream and to distortion of the large eddies as they travel over the hill. Distortion of the mean flow affects the turbulent structure when the air's travel time over a hill is short compared to the time scale of that structure, as discussed by Britter et al. (1980). In this case, rapid vertical compression and horizontal stretching of large vortex elements would decrease the energy of the longitudinal component and would increase the energy of the lateral and vertical components.

The spectrum at 86 m is in the outer region, where little production of mechanical turbulence occurs and where diffusion of turbulent

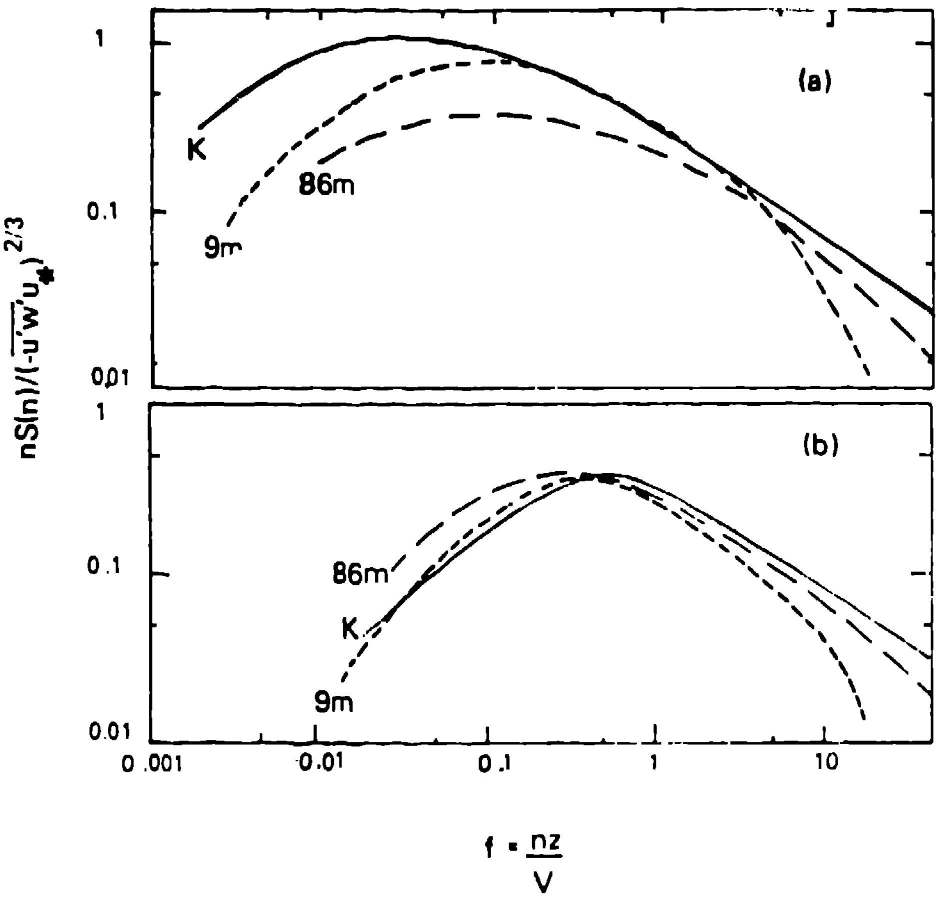

Fig. 4. Normalized spectra for the u component (a) and the w-component (b) in near-neutral conditions at 9 m and 86 m above the crest of Black Mountain. The solid line is the flat-terrain model suggested by Kaimal et al. (1972).

energy is expected to play an important role. The assumptions in Eqs. (7) and (8) are not valid in this region, so scaling according to Eq. (9) is inappropriate. However, the shape of the spectrum is qualitatively what one would expect; that is, a shift of the peak to higher frequencies as lower frequencies lose energy and high frequencies adjust to a larger surface stress.

Vertical-velocity spectra are shown in Fig. 4b. The lack of significant difference between the observed spectra and the flat-terrain model is explained by the fact that vertical-velocity spectra span a higher frequency band than do horizontal-velocity spectra and, therefore, will be in quasi-equilibrium with underlying terrain. Both memory and distortion will affect the very large eddies, those associated with $f < 0.05$, but in the vertical-wind component these two effects tend to cancel one another; this may explain the good agreement between the model and the observed spectra at very low frequencies.

Bowen (1979a and b) measured wind characteristics on two escarpments in New Zealand. One has a vertical face with a height h = 11.6 m; the other has a slope of 1 (vertical) to 2 (horizontal) and a height h = 13 m. Towers at each site were located at x = -10 h, 0.5 h, and 4 h, where x is the distance downwind of the escarpment edge.

Because of the short fetch up these escarpments, the inner region is expected to very very thin, probably 1 m for the sloping escarpment and essentially zero for the cliff. Hence, all observations were made in the outer region.

Figure 5 compares unnormalized spectra of the longitudinal-velocity component measured at a height of 1.6 m at the three towers over the sloping escarpment. The reduced energy at very low frequencies may be due to rapid distortion of the mean flow; however, we would have expected this reduction to extend to much higher frequencies. The high-frequency energy remains essentially unchanged, even though the wind, and presumably u_*, increase by about 30%. The reason for this result is that, even for the shorter wavelengths, eddy-response time is long compared to travel time up the escarpment, so these eddies are not in local equilibrium.

Figure 6 shows spectra measured over the cliff escarpment. The change in shape of the spectrum between x = -10 h and x = 0.5 h is quite spectacular. Because the flow separated in this region, we shall not attempt to apply our current theory to explain these spectra.

Finally, we consider a much larger escarpment within the White Sands Missile Range, New Mexico. Vertical terrain cross sections from the south-southwest and from the east are shown in Fig. 7; relatively reliable data were available when winds blew from these directions. Unfortunately, only 10-s averages were recorded, so that only the low-frequency portions of the spectra could be obtained.

Upwind spectra are not available, so again we used the Kaimal model to approximate the spectrum for undisturbed flow. We used the neutral limit of the stable-air model because observations were made at night during strong winds.

The normalization scheme used for the Black Mountain spectra is used here. An apparent u_* value was computed from a logarithmic wind profile fitted to wind speeds at 8 m and 16 m. The u_* values were also estimated by extrapolating the local Reynolds stress ($\overline{-u'w'}$) profile; a cospectral model (Panofsky and Mares 1968) was used to correct measured local stress for high-frequency losses. The agreement between these two results for u_* was quite good.

Figure 8a shows spectra for winds from the south-southwest. The slope in this direction is very steep, and the fetch over the last major change in slope is about 200 m. According to our second proposition in Section I, longitudinal spectral densities should be affected by a lower upwind surface stress and by distortion of the mean flow. Both effects contribute to an energy deficit, and this is what we observe. And, because vertical compression of the mean flow tends to increase the turbulent energy of the lateral-wind component, the observed energy deficit should be less than in the longitudinal component. Again, this is what we see in Fig. 8a.

Because the fetch for easterly flow is of order 2000 m and the slope is gentle, even the largest eddies have had time to adjust to upslope conditions. Spectra for this wind direction, shown in Fig. 8b, resemble equilibrium flat-terrain spectra, in good agreement with our hypothesis.

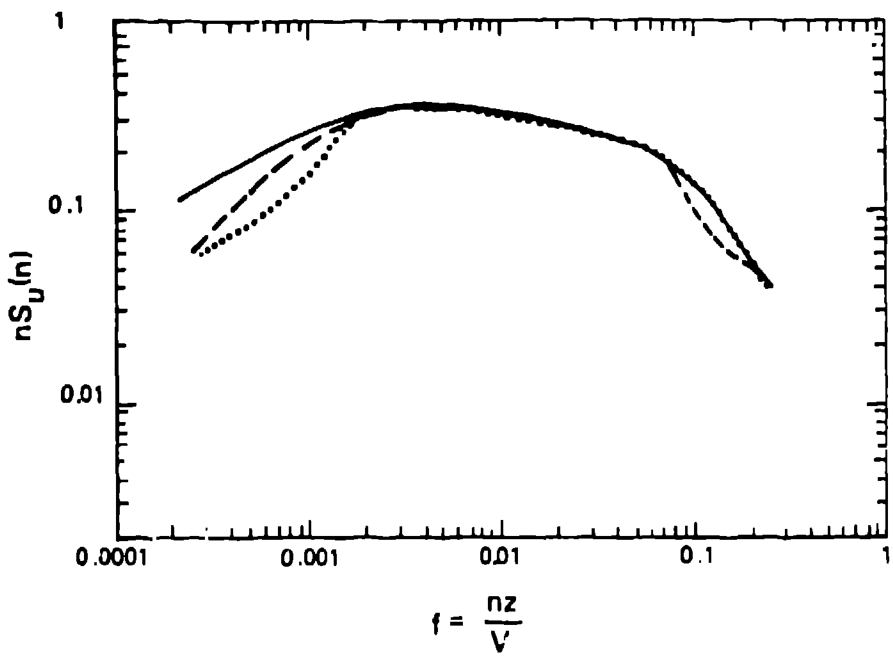

Fig. 5. u spectra in near-neutral conditions at a height of 1.6 m above the sloping escarpment, New Zealand. x = -10 h (solid line), x = 0.5 h (dashed line), x = 4 h (dotted line).

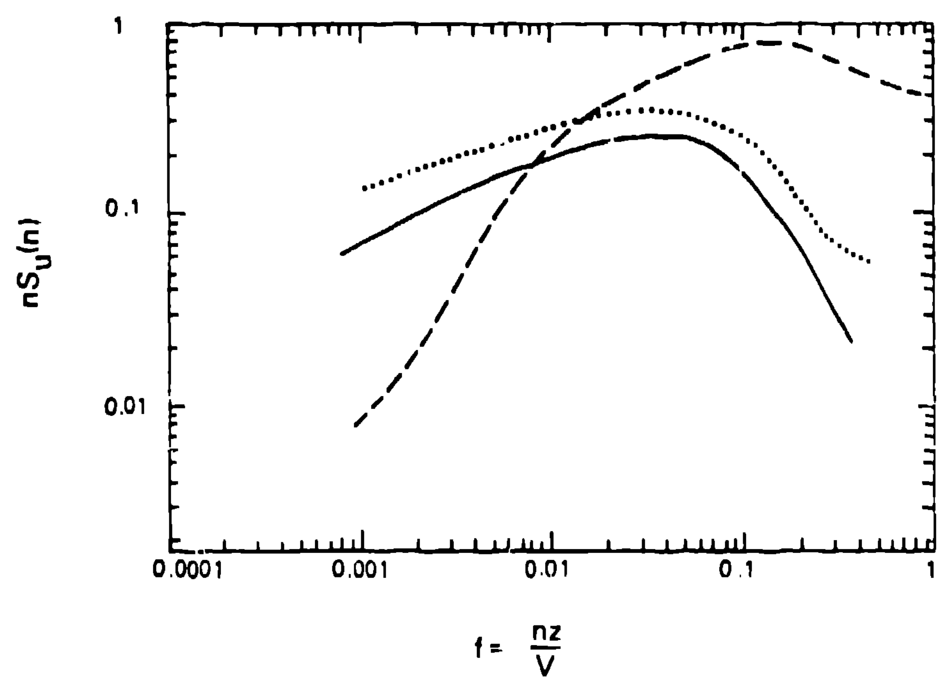

Fig. 6. u spectra at a height of 1.6 m above the cliff escarpment, New Zealand. Details as in Fig. 5.

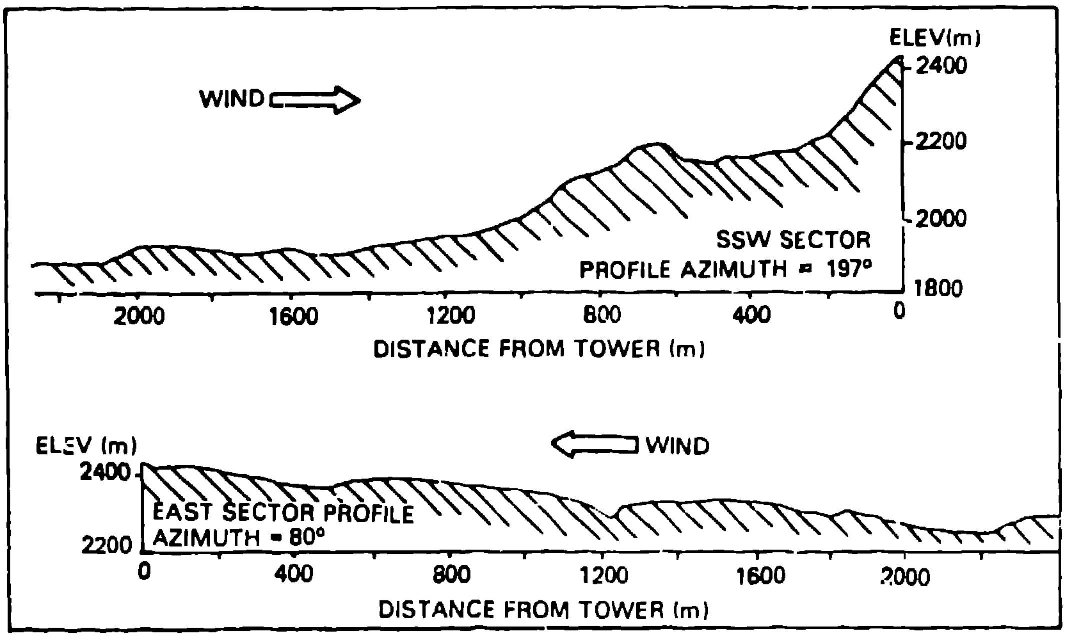

Fig. 7. Terrain cross sections at White Sands.

IV. SPECTRA OVER ROLLING TERRAIN

Spectra of all velocity components were obtained at a rural site at Rock Springs, Pennsylvania (near State College), by members of the Meteorology Department, The Pennsylvania State University. Observations were made at 5 m and 8 m above locally flat terrain, with roughness length of order 1 cm, surrounded by rolling farmland. South and southeast of the site there is a nearly two-dimensional wooded ridge that rises about 200 m above the surrounding plain. The site has been described by Perry et al. (1978).

Observations were made in stable and unstable conditions and at levels where the local Reynolds stress does not differ significantly from u_*, so the spectra were normalized by $u_*^2 \phi_\epsilon^{2/3}$, as is customary over uniform terrain (Kaimal et al. 1972). When normalization includes the dimensionless dissipation ϕ_ϵ, spectra become independent of stability at high frequencies, where the spectrum function is again described by $af^{-2/3}$, as in Eq. (9).

For these observations, u_*^2 was estimated from the u spectra by taking $nS(n)/u_*^2 \phi_\epsilon^{2/3} = 0.3$ at $f = 1$. The value of ϕ_ϵ in stable air, according to Kaimal et al. (1972), is given by

$$\phi_\epsilon = [1 + 2.5(z/L)^{3/5}]^{3/2} , \qquad (10)$$

where L is the Monin-Obukhov stability length. In unstable air ϕ_ϵ is given by

$$\phi_\epsilon = (1 - z/L) , \qquad (11)$$

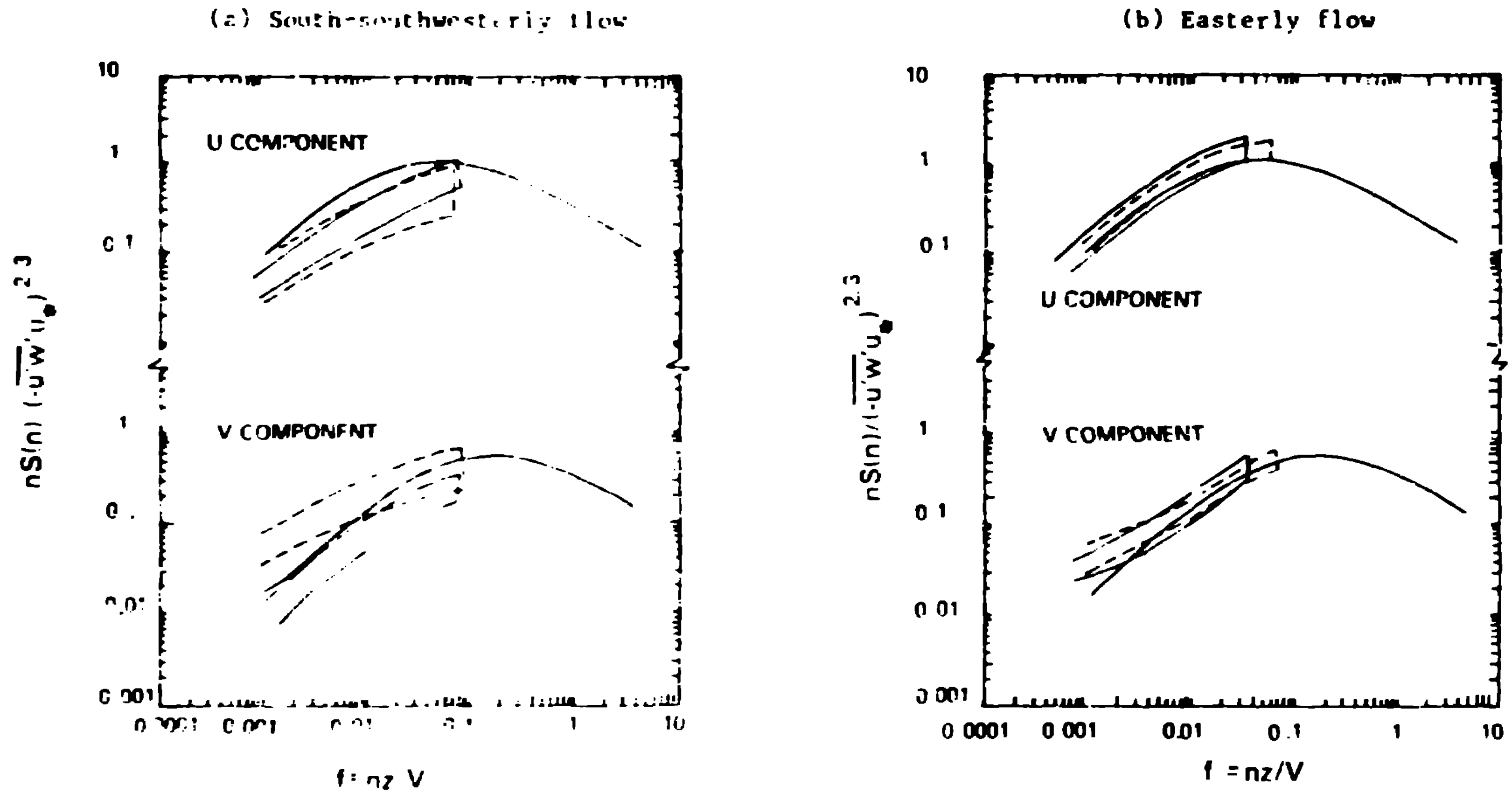

Fig. 8. Envelopes for normalized spectra in near-neutral conditions for south-southwesterly flow (a) and easterly flow (b) at White Sands. Solid envelope for $z = 8$ m, dashed for $z = 16$ m. Smooth curves are flat-terrain models suggested by Kaimal et al. (1972).

which is a compromise derived from Kaimal et al. (1972 and 1976) and unpublished observations of Kaimal at the Boulder Atmospheric Observatory.

Figure 9 shows envelopes of spectra of the velocity components in stable air that has not crossed the ridge. Models developed by Dutton et al. (1979) for flat terrain, differing only slightly from Kaimal's model, are shown for comparison. The model curve is based on the average value of z/L in each sample. Model and observations have been matched at high frequencies on the premise that, in that regime, the spectral densities have adjusted to local terrain.

Only for the horizontal components under stable conditions is there evidence of departure from the flat-terrain model. This departure is a very marked increase in low-frequency energy, in contrast to the relative reduction at this end of the spectrum that attends rapid adjustment to a change in roughness (from smooth to rough) or flow up a steep hill. As before, the w spectrum appears to be in local equilibrium at all frequencies.

This increase is consistent with the observation of Panofsky et al. (1979) that σ_u/u_* at the same site is larger than expected over flat terrain. This may be due to the "memory", in the low frequencies, of the larger roughness or more hilly terrain upstream. Because the increase occurs in stable conditions, the enhanced low frequencies may also reflect the presence of gravity waves or slow horizontal meandering of the flow around terrain irregularities. This possibility is supported by spectra of the two horizontal components in stable air downwind of the ridge (see Fig. 10). The low-frequency excess [illegible] is even larger than for air with trajectories over the plain.

Finally, Fig. 11 shows spectra of all velocity components in very unstable air ($z/L < -0.1$). For each component, the spectra fit the uniform terrain model at all frequencies. The inference is that terrain features are relatively unimportant in very unstable air.

V. CONCLUSIONS

In general, when neutrally stratified air flows over [illegible] terrain, the spectrum of the turbulent structure is continuously [illegible]fied. The high-frequency structure adjusts rapidly to new [illegible] conditions. Low frequencies respond more slowly, and when streamlines are rapidly distorted, there is evidence that turbulent energy [illegible] redistributed among the velocity components. Specifically, we [illegible] the following:

- Within the internal boundary layer downwind of an increase of roughness, high-frequency spectral densities are [illegible] than they were upstream. Energy of the low frequencies [illegible] unaffected 2000 roughness lengths downstream.

- When spectra obtained in the inner region over hilltops are scaled by [illegible] (on the basis of equilibrium between local production and dissipation rates), close conformity to a flat-terrain model spectrum is achieved in the inertial subrange.

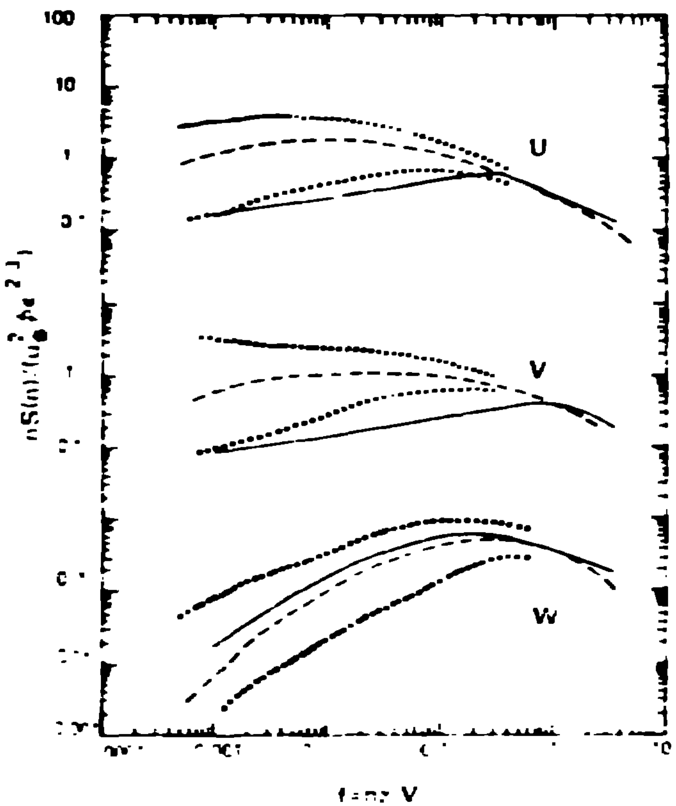

Fig. 9. Normalized average spectra for the three wind components in stable condition at Rock Springs. *Dashed lines represent the* means, dots the extremes. The solid lines are flat-terrain models suggested by Dutton et al. (19[illegible]).

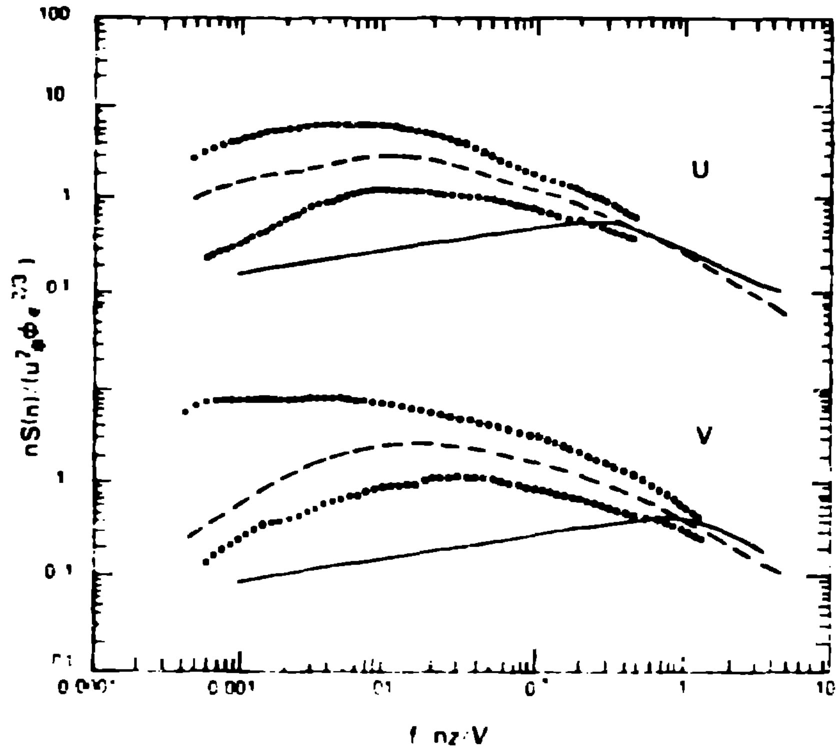

Fig. 10. Normalized average spectra for mountain trajectories in stable conditions at Rock Springs. Details as in Fig. 9.

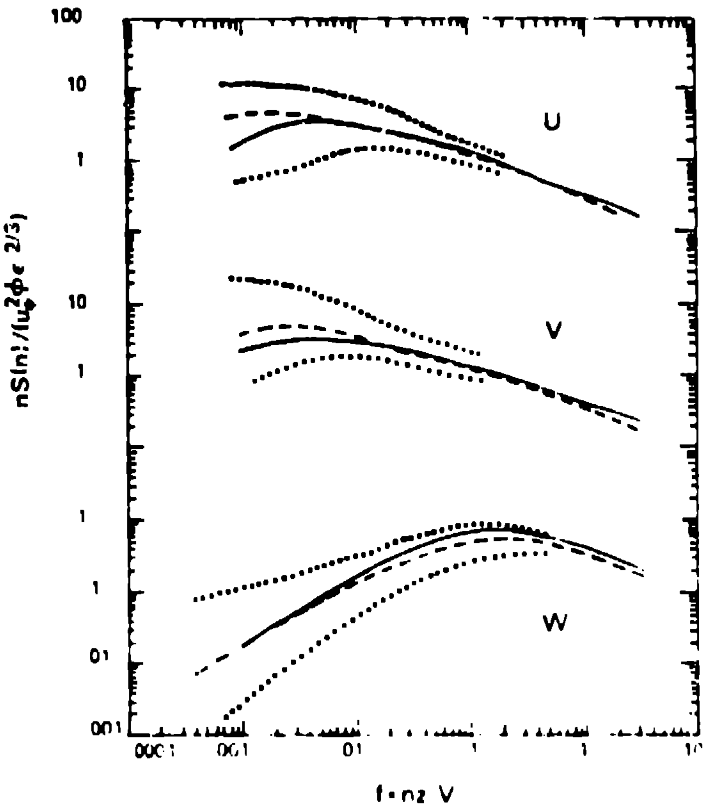

Fig. 11. Normalized average spectra for the three wind components in unstable conditions at Rock Springs. Details as in Fig. 9.

- Normalized u spectra obtained over hilltops show a pronounced energy "deficit" at low frequencies. This deficit can be explained by the long response time of large eddies, which allow them to remember conditions upstream and to respond to distortion of the mean flow.

- Vertical-velocity spectra obtained over complex terrain resemble those obtained over flat terrain. This resemblance is explained by the fact that energetic vertical-velocity fluctuations occur at higher frequencies than do horizontal-velocity fluctuations and, therefore, are in quasi-equilibrium with local conditions.

- Velocity spectra obtained over rolling terrain in unstable conditions resemble flat-terrain spectra. In stable conditions, spectra of the horizontal-velocity components show considerably more low-frequency energy than is observed over flat terrain.

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