SOME OBSERVATIONS OF AIRFLOW OVER A LARGE HILL OF MODERATE SLOPE

M. W. GALLAGHER, T. W. CHOULARTON, and M. K. HILL

Department of Pure and Applied Physics, U.M.I.S.T., Sackville St., P.O. Box 88, Manchester M60 1QD, U.K.

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Abstract. Measurements are presented of mean windspeed and turbulence over Great Dun Fell, which is rather larger than hills investigated in the past, viz., 847 m high, which is comparable to the boundary-layer depth. The Fell is well suited for study, being covered by rough grass with no trees and few other obstructions. It was found that the speed-up of the wind is dominated by the elevated stratification and generally agrees closely with the predictions of the model of Carruthers and Choularton (1982) except when the flow is blocked. On the hill summit, the turbulence is approximately in local equilibrium in at least the lowest 10 m and the turbulence measurements are similar to those obtained within the inner layer at other sites. The transverse and longitudinal components show spectral lags at wavelengths greater than 30 m. This suggests an inner-layer depth of about $\frac{1}{3}$ that predicted by Jackson and Hunt (1975). At reduced frequencies (<0.1), a recovery in spectral energy is observed due to gravity wave activity. A large variation in the streamline tilt at the summit is observed depending on whether the airflow regime is supercritical or subcritical.

1. Introduction

The problem of how variations in topography and roughness affect turbulent flow structure has received increasing attention recently due to its importance in modelling the turbulent boundary layer, pollution transport, and wind energy conversion systems.

Most workers have focused on the 'flow over a hill' theory of Jackson and Hunt (1975, hereafter referred to as JH). Since it has the advantage of highlighting the flow dynamics, this model has contributed considerably to the understanding of how wind flows over a variety of terrain types where a change in slope or roughness or both occur.

The theoretical predictions concerning the mean speed-up over hills are borne out by observations; for example, see the extensive review by Hunt (1980). The general acceptance of the JH model is in spite of sparse experimental evidence to support the more detailed predictions of the theory (Britter *et al.*, 1981).

Recently, Mason (1986) has presented measurements made on a roughly circular hill 70 m high and with a base diameter of 500 m. He finds that whilst the mean speed-up of the wind is consistent with the predictions of the generalised JH theory, the small-scale turbulence is only in local equilibrium in the lowest fraction of a metre above the surface at the hill summit. He argues that the inner-layer depth is substantially less than that predicted by JH.

Large-scale turbulence, on the other hand, adjusts quite slowly to changes in terrain. Near the surface of a hill, in the inner layer, equilibrium between mechanical shear production and dissipation of turbulent kinetic energy exists since the transit time for these eddies over the hill is large compared to their characteristic mixing times. The turbulence levels within this layer will then be modified as indicated by Britter *et al.* (1981).

In the outer layer where the large-scale eddies dominate the flow, there is no balance between shear and dissipation. The eddies are advected relatively quickly in comparison with their turnover times without losing energy but suffer distortion by the mean shear of the flow. Significant changes in the turbulent structure occur as the eddies are effectively stretched. Britter *et al.* estimate the subsequent perturbations in the turbulence levels using rapid distortion theory and show that there is a significant increase in the variance of the vertical and transverse components compared with a decrease in the longitudinal component. This behaviour was observed by Bradley (1980) on Black Mountain; however, no decrease in the longitudinal component was observed and the increase in the vertical component was larger than for the transverse. This vortex stretching mechanism would cause an apparent deficit in the longitudinal spectrum for scales greater than the inner-layer height.

An appropriate normalisation for such spectra was suggested by Panofsky *et al.* (1982), to be $(-u'w'u_*)^{2/3}$, where $-u'w'$ is the local stress and u_* the local friction speed, derived from profile measurements. This normalisation was confirmed by Smedman and Bergstrom (1984). A more correct terminology describing the difference in response between the two spectral regions is 'spectral lag' introduced by Högström *et al.* (1982).

The need for a clearly defined approach flow is shown by the differing behaviour observed by a number of workers. Although Panofsky *et al.* (1982) observed a spectral deficit or lag for scales greater than the inner layer height at Black Mountain, Smedman and Bergstrom (1984) observed an enhancement at low frequencies over a very low hill as did Neal (1981) at the crest of a steep ridge. Measurements by Högström and Smedman-Högström (1984) over complex, forested escarpments revealed substantial lag effects, but the departure from the model of Kaimal *et al.* (1972) occurred for scales not apparently consistent with the JH predicted inner-layer heights (mostly confirmed by wind profile measurements).

The nature of the spectral modification especially under stable conditions, where buoyancy forces are important together with a possible wave contribution, is still under investigation, particularly for large-scale hills. Bradley (1983) presented observations of flow over a very low, smooth, unsymmetrical hill under a variety of stability conditions. The predicted JH inner-layer depth of 3 m did not reveal itself in the profile measurements and no jet was observed. Despite this, the speed-up agreed well with that predicted by theory for unstable conditions. The discrepancy between observed and predicted speed-ups for stable stratification was later resolved by Hunt and Richards (1984), who introduced the effects of upper-layer buoyancy forces.

2. The Great Dun Fell Site

The Great Dun Fell site (GDF), shown in Figure 1a, has been described in, for example, Baker *et al.* (1982). Essentially GDF is a large nearly two-dimensional ridge situated

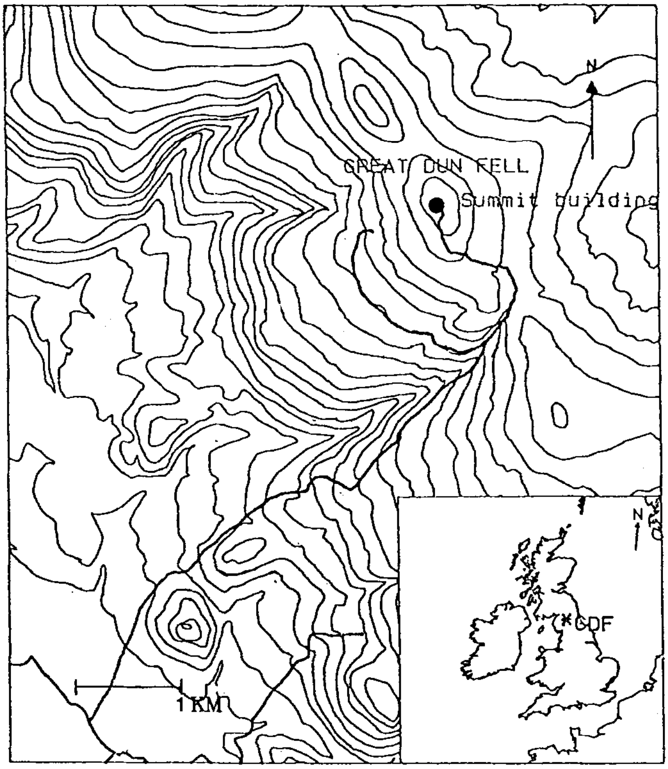

Fig. 1a. A 15 m contour map of Great Dun Fell showing the summit and the surrounding terrain.

in the Northern Pennine ridge, which runs approximately NW-SE, overlooking the Eden Valley to the SW. It is the second highest peak in the Pennines at 847 m a.s.l. and is also the site of the highest Meteorological Office Synoptic Station in the U.K. Using the hill nomenclature of Hunt (1980), the height of the hill above the surrounding valley plain is h = 665 m while L, the length at half-height, is 2 km. The site is very exposed with no trees and consists mainly of grasses about 10 cm in height.

In addition for about 250 days of the year, the summit of GDF is in cloud enabling a variety of cloud microphysical and chemical measurements to be made. This site provides an ideal laboratory for airflow and cloud interaction experiments. The Atmospheric Physics Research Group (APRG) also maintains an instrumented complex within the valley some 7 km south of GDF summit so that orographic effects can be investigated. The measurements to be presented here form a small part of these much larger experiments generally concerned with in-cloud processes. The present measurements were for periods when the summit of the hill was out of cloud.

The mean roughness length as measured by a number of workers at the summit of GDF is $\simeq$ 2–4 cm. More recent measurements have confirmed this value at a number of other locations on the hill; the value is consistent with the values quoted in the literature for this type of terrain. Recent construction work required landscaping to be carried out at the summit with the result that in places the grass is very short. Roughness lengths over these fetches of re-establishing grass were found to be of $\simeq$ 2–4 mm. Using these terrain scales, an inner-layer depth can be calculated from Jackson and Hunt,

$$l \ln (l/z_0) = 2k^2 L \,, \tag{1}$$

where l is the inner layer depth, z_0 the roughness length, L the length at half-height, and k von Kármáns's constant $\simeq 0.4$. This yields a value of l for GDF of $\simeq 80$ m. The recent work of Mason (1986) suggests that a closer approximation to l is given by

$$l(\ln (l/z_0))^2 = 2k^2 L \,, \tag{2}$$

which yields a value of $l \simeq 14.7$–16.9 m.

3. Data Acquisition

The analogue data from the ultra-sonic and cup anemometers together with a variety of other meteorological parameters were logged using an inexpensive portable data acquisition system, designed within the APRG. This system is extremely flexible and can be configured to meet the needs of any number of instruments, simultaneously, that may be on site. It is a Distributed Real Time Input Processing System (DRIPS) built around the Z-80 based 64 K Nascom 2 microcomputer. The data is logged at 40 Hz for sonic and at 4 Hz for general meteorological parameters via 12-bit analogue-to-digital converters. It is then stored on industry standard $\frac{1}{2}''$ magnetic tape using a Pertec 9 Track, 1600 b.p.i. tape drive. The Nascom microcomputer was also used for the subsequent analysis of the data.

4. Site Considerations

Figure 1b is a 1-m contour map of the summit of GDF (the zero datum taken as the highest point). The instrumented complexes at the site are approximately 3 m high. The feature labelled 'B' is the instrumented platform on which various instrumentation is mounted. The turbulence measurements considered here were made using a Kaijo-Denki DAT-300 ultra-sonic anemometer with a TR61–A probe. The platform is approximately 3 m high and the anemometer was mounted on a tripod arrangement fixed to the platform giving a maximum measurement level of 7.8 m above the ground. Cup anemometers at 5 and 10 m were also available for comparison with the sonic anemometer measurements, although these were not designed for accurate profile measurements. An immediate concern was the possible influence of nearby structures on the measurements; however, the variation of this influence with wind direction indicates that for the prevailing wind directions encountered, these are negligible. Sonic-derived U_* and profile-derived velues for steady wind conditions agreed to within 5%. Data sets where the wind direction moved beyond an acceptable angle from the ideal were rejected since large perturbations in the local stress occurred. Additional

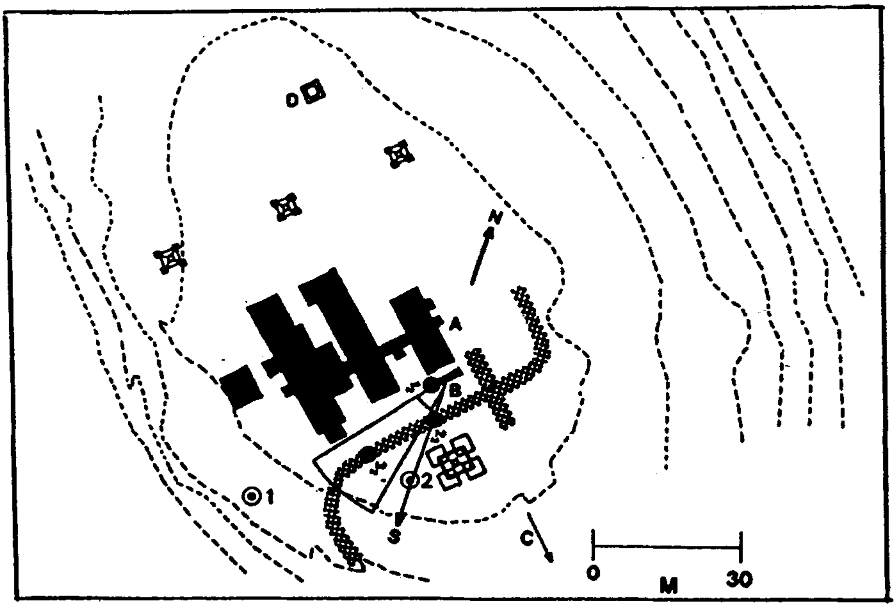

Fig. 1b. The UMIST APRG instrumented complex, marked '*A*', at the summit of Great Dun Fell. The contours are in 1 m intervals referred to the summit plane. The general acceptance angle for measurements is shown. '*B*' marks the position of the instrumented platform and 1 the position of the summit lip. '*D*' marks the position of the Meteorological Office 10 m instrumented tower.

problems encountered at this site arose from the frequent periods of cloud and rain. Such periods tended to contaminate the sonic data to unaccceptable levels and were also rejected.

The influence of the platform mount on the stress measurements is more difficult to determine. The sonic was aligned using bubble levels at the top and base of the tripod mast. The flux term was transformed to the local vertical using the usual algebraic tilt corrections, e.g., Pond (1968), Kaimal and Haugen (1969), and Tanner and Thurtell (1969). Quite substantial streamline tilt can occur at this site depending upon whether the flow is supercritical or subcritical. Generally the mean tilt angle was close to zero, and the corrections were small; however, a maximum of 6 deg was encountered during some case studies for NE wind directions, and the subsequent corrections were of the order of 30%. Such tilt angles can lead to large errors in flux measurements, the sensitivity being such that under changing flow conditions, complete reversal of sign of $u'w'$ can occur over complex terrain, e.g., Mason and King (1984). Measurements such as those of Hyson *et al.* (1977) indicate that the tilt corrections are of use in correcting errors due to misalignment of the probe or tilted flow cases; however, they are not applicable to correction for possible distortion to stress by, e.g., probe mounts. Wyngaard (1981) shows that the tilt correction may only account for about half the error induced in the stress measured in such cases. That the induced errors cannot be simply corrected for using the tilt equations is shown by the measurements of Coppin and

Taylor (1983). In general for the unperturbed flow cases, we find our sonic measurements underestimate stress values derived from 5 and 10 m cup anemometers by about 10–30% at most, depending on wind conditions. This error could also include a doubt about the calibration and overspeeding of the cup anemometers.

The platform itself is situated some 40 m from the first SW contour in Figure 1b, where a small but sharp lip occurs. Because of the concern over possible intermittent separation effects from this lip, a series of measurements were made using the sonic mounted at a level of 3 m at several sites at and downwind of the lip. These sites are labelled 1, 1_2, 2_2, 3_2, 2, and 3 in Figure 1b. They showed that there was a large vertical wind component near the edge. The turbulence field here did not appear to be in equilibrium, with very small values of local stress recorded. The high frequency spectra showed higher spectral levels than the Kaimal model prediction, being very similar to the escarpment spectra observed by Bowen and Lindley (Panofsky *et al.*, 1982), near

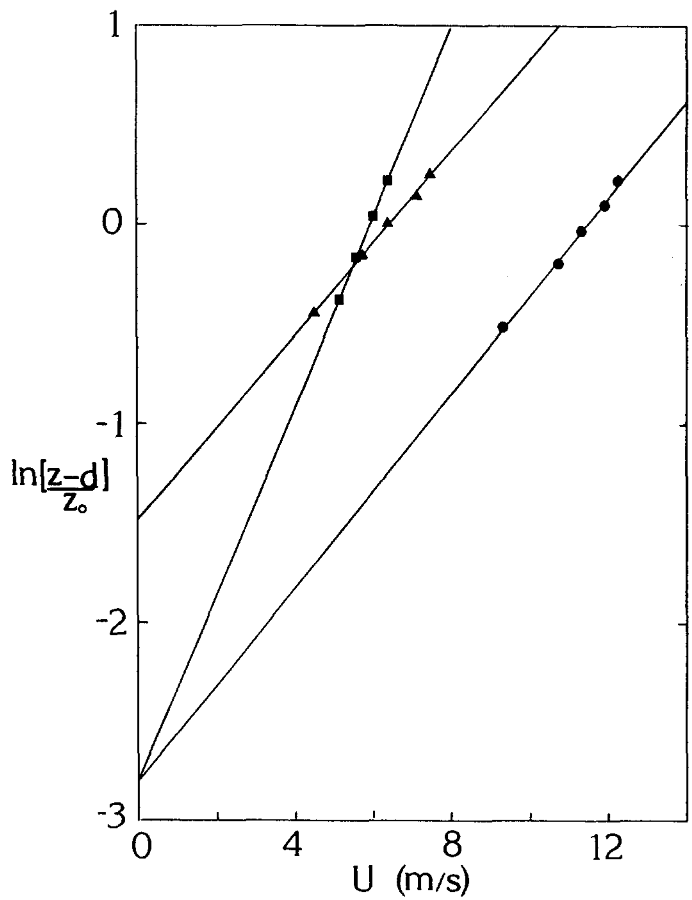

Fig. 2. Anemometer profiles obtained at site 2 (●) and site '*D*' (■) at the summit, $Z_0 \simeq 2$ mm and 172 m below the summit (▲), $Z_0 \simeq 33$ mm. d is a zero-plane displacement $\simeq \frac{2}{3}$ the height of the roughness elements.

a large slope change. The 3 m measurements downwind of site 1 did not show any significant evidence of eddy shedding, indicating that the resulting internal boundary layer is below this level. The 7.8 m level measurements appear to show little or no perturbation as a result of this lip effect, although the turbulence intensity levels are higher at this site than found over flat terrain, as discussed later. It should be stressed that the turbulence measurements made on our platform showed very similar features to the 3 m measurements made at other sites on the summit for shorter periods.

Figure 2 shows a typical series of profile measurements made at two locations on the summit for different wind speeds, and at a location upwind some 172 m below the summit level. These measurements were made with cup anemometers mounted on a 3 m tower. The anemometers had been carefully calibrated in a wind tunnel prior to the experiment.

The two profiles measured at the summit were obtained on either side of the instrument complex, one close to the platform site and the other near the tower marked 'D' in Figure 1b. The summit measurements were made on different days with wind speeds of 8 and 12 m s^{-1}, respectively, and for SW winds. Good logarithmic profiles were found and agreement between the two sites was good, indicating local equilibrium in the lowest 3 m. The roughness length was very small, $\simeq$ 2 mm because the grass cover had yet to re-establish itself after recent construction work. The profiles obtained further down the SW slope were for more normal grass cover and yielded roughness lengths $\simeq$ 3 cm. These measurements lend confidence to the validity of the measurements from the sonic anemometer. In addition, measurements were made using hand-held anemometers at a large number of points on the summit. These confirmed that the mean wind speeds measured at the fixed anemometers and by the sonic anemometer were not affected by local terrain features or by the buildings in a SW flow.

5. Results and Model Comparisons

5.1. Synoptic Situation

The following case study data were observed primarily within a period of five days, 23 to the 27 May, 1985, at Great Dun Fell during which several experiments were being conducted.

During this period, a generally SW airflow covered the area with depressions moving NE to the NW of Scotland. A cold front crossed the area on the 24th and a warm front on the 25th; pressure built on the 26th with subsidence in the middle troposphere.

These changes resulted in marked variations in the stability structure of the troposphere which would be expected to produce substantial changes in the airflow pattern over GDF. Hence, the period provided an opportunity to test the ability of the model. Figure 3 shows the mean wind speeds measured 10 m above the local terrain for GDF, Wharleycroft and also Shap, which is about 30 km further upwind.

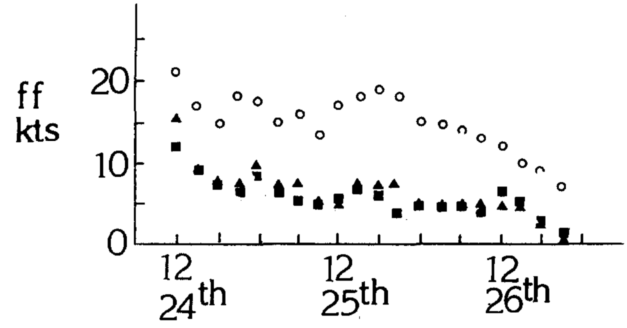

Fig. 3. Mean 10 m wind speeds observed at GDF, ○, Wharleycroft, ▲, and Shap, ■, for the period 24–27 May, 1985.

5.2. Comparison of the Observed and Predicted Speed-up of the Wind

Mean wind directions recorded at the summit of Great Dun Fell and Wharleycroft were in good agreement, although the upwind site exhibited greater variation due to the tendency for blocking of the flow.

The airflow model of Carruthers and Choularton (1982) was used to compare predicted speed-ups with observations. This extends the Jackson–Hunt theory to include the effects of upper-level stability. The basic features of the model are outlined in Figure 4. Three layers are defined to describe the structure of the boundary-layer flow of an elevated inversion over the hill. The lowest layer, L1, is assumed to be neutral; the second layer, L2, is an inversion capping L1; while the third layer, L3 extends from the inversion top. A stability parameter for each layer is defined using the equation,

$$\mu_i = ((1/\theta)\delta\theta/\delta z(1/U^2))^{1/2}, \tag{3}$$

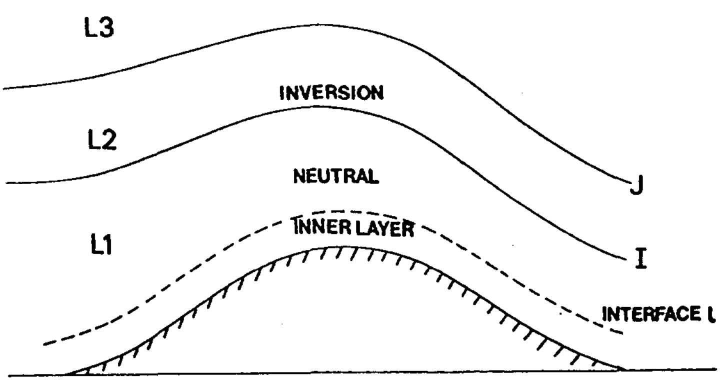

Fig. 4. A diagram showing the main features of the Carruthers and Choularton (1982) three-layer airflow model. L1 is the neutral lower layer, L2 is the capping inversion, and L3 is a layer extending to the lower troposphere. The dashed line marks the inner layer. I and J mark the interface heights above the valley floor.

where μ_i is the layer stability ($i = 1 \ldots 3$), θ the mean potential temperature in the layer, z the thickness of the layer, and U the mean wind speed in the layer. These input stability parameters were calculated from 12-hrly sonde ascents from Aughton, located about 100 km SW of Great Dun Fell.

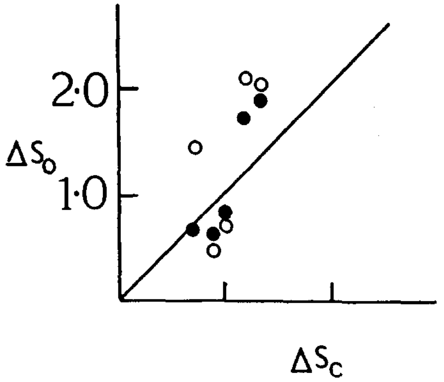

Fig. 5a. A scatter plot of the observed ΔS_0 and model calculated ΔS_c speed-up ratios for Great Dun Fell during the period 24 to 26 May, 1985.

Figure 5a is a scatter plot of model predicted and observed speed-up ratios. The greatest deviations between calculated and observed speed-up ratios occur for noctural periods where stable stratification prevails in the lower layer. This can lead to blocking of the flow upwind of the hill as the nocturnal inversion forms with little or no

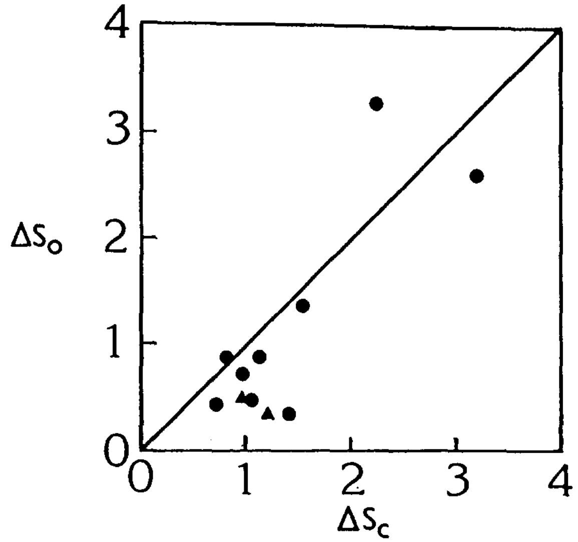

Fig. 5b. A scatter plot of observed, ΔS_0, and calculated, ΔS_c, speed-up ratios for Great Dun Fell for a longer period prior to the case study, with near constant wind directions between the hill summit and the upwind station.

perturbation to the higher level airstream. Two out of the three daytime periods show good agreement. In general, when the flow was not blocked, the predicted values of the speed-up were somewhat greater than the observed ones.

This observation is in agreement with calculations made for another data set selected with near constant SW wind direction at both Wharleycroft and Great Dun Fell, thus eliminating blocking conditions. These are shown in Figure 5b. The correlation between observed and calculated values is 0.9. Model predictions are consistently greater than observed values. This is probably due to the three-dimensional nature of the flow over Great Dun Fell, where lateral flow divergence will be important, reducing the overall value of the speed-up ratio. Nevertheless, the results clearly show that the model is able to make useful predictions of the variation in speed-up as atmospheric conditions change.

5.3. Interpretation of Speed-up Changes during the Experiment

Figure 6 shows the temporal variation of fractional speed-up ratio, ΔS, calculated using both the Wharleycroft and Shap sites as the upwind reference sites compared to model predictions. The agreement between speed-ups calculated from the two upwind sites is reasonable; occasionally, e.g., in the regions marked *A* and *B*, a tendency for blocking of the flow at the Wharleycroft, close to the base of the hill occurs under nocturnal conditions. This gives a much higher apparent speed-up from that site.

Considering the flow regimes in more detail, the period 1200 LST on the 24th to 0000 LST on the 25th, was one of high geostrophic wind speed, and a significant positive streamline tilt was observed at the summit of Great Dun Fell, $\simeq 1.5$–2.0 deg, suggesting separation of the flow in the lee of the hill. The fractional speed-up ratio was close to

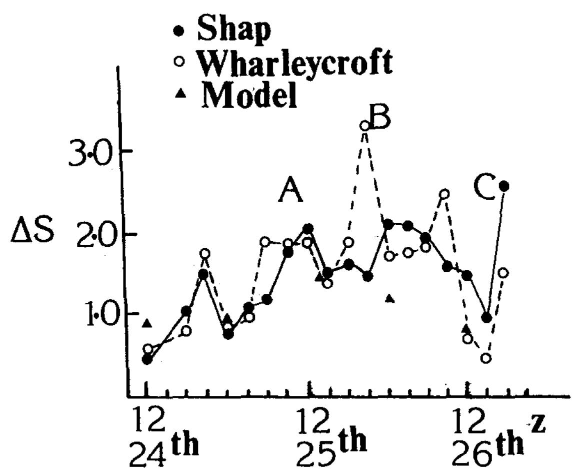

Fig. 6. The temporal variation of speed-up ratio calculated for Great Dun Fell using upwind data from both Shap and Wharleycroft. The model predictions are also shown for comparison. The period marked '*A*' is mostly stable with subcritical flow, '*B*' represents the onset of blocked flow in the Eden Valley, and '*C*' again shows a return to blocked flow.

the neutral value initially. As the geostrophic wind speed decreased, the flow showed a transition from neutral or slightly supercritical flow to a subcritical regime with an increase in speed-up over the neutral value. The vertical wind angle at Great Dun Fell summit was observed to be slightly negative. A sudden decrease in ΔS was observed at 0000 LST on the 25th, indicating a return to near neutral or slightly supercritical flow, with the vertical wind angle again close to zero or slightly negative. This period was one where the flow was essentially dominated by changes in elevated stratification and the speed-ups calculated from the Wharleycroft and Shap sites agree quite well.

The following period was dominated to a large extent by blocking, with the flow becoming markedly subcritical. The sudden onset of blocking was evidenced by a significant decrease in wind-speed at Wharleycroft. There was also evidence for this at Shap a little later. A clear indication of blocked flow in the Eden valley occurred towards the evening on the 25th with reasonable agreement between the two upwind sites, but with markedly different behaviour from the airflow model predictions, the observed speed-ups being much greater than the calculated values. The model of Carruthers and Choularton (1982) cannot reproduce the effects of blocking. The return to non-blocked flow during the early morning of the 26th is marked. Radiosonde data for this period showed the formation of a low-level subsidence inversion, and the flow thus became supercritical, with very low speed-ups observed and predicted. After 1300 LST on the 26th, the winds became very light with blocked flow at both upwind sites, point C.

A comparison between the local stability observed at the hill summit and the associated speed-up was performed. No clear relationship was found in contrast to Bradley's (1983) results for a very small hill. This is because over the larger Great Dun Fell, the value of ΔS is controlled by the profile of stratification throughout the lower troposphere.

5.4. Summit Turbulence Measurements

Since turbulence measurements were only available from the single sonic anemometer, no profile information from simultaneous measurements can be presented. The measurements were restricted to the maximum level of 7.8 m above the surface. Separate measurements close to the summit edge revealed the presence of a 'lip' effect, similar to that observed by Bowen and Lindley over small escarpments (Panofsky *et al.*, 1982). Stress measurements close to this edge, 3 m above the surface, were $\simeq 3$ times smaller than observed at the same height close o the platform. A relatively large wind angle of $\simeq 7$ deg was also observed at the edge compared to a near-zero value by the platform. Stress measurements made at 3 m by the platform and at 7.8 m on the platform showed good agreement.

Data from the sonic were collected from several periods, mostly near-neutral, but with some very unstable periods. Statistical parameters of variance and stress were averaged over 15 min intervals. To reduce scatter, the horizontal components were placed into stability bands and averaged. The vertical component, because of its short characteristic response time, $\simeq 10$–30 s, showed very consistent behaviour for smaller averaging times. Figure 1 shows the variation of σ_w normalised with the local stress-derived U_*,

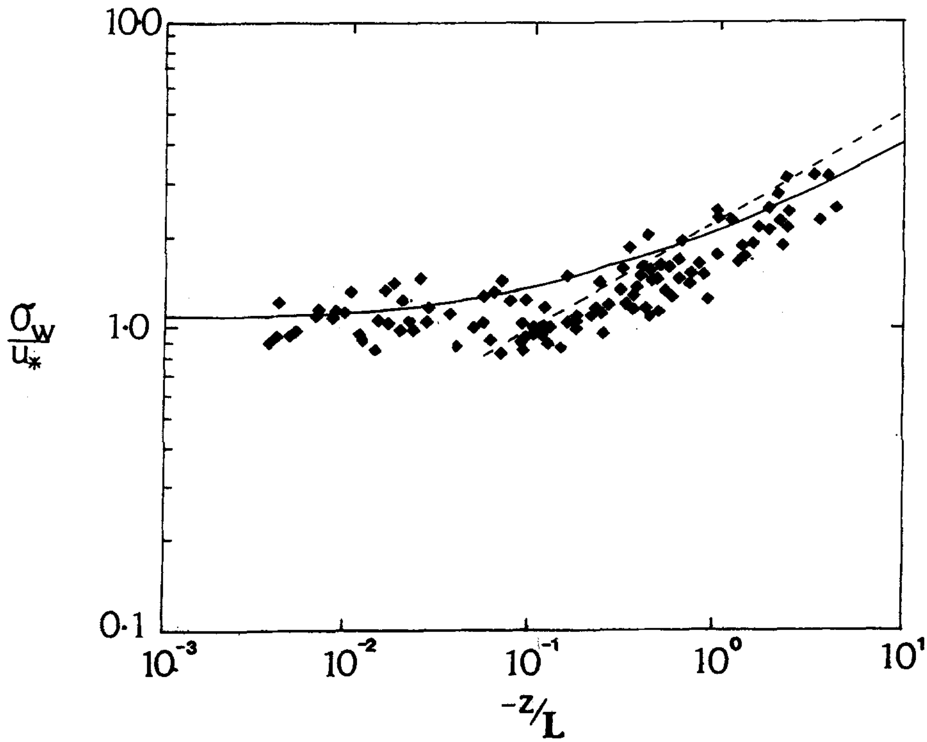

Fig. 7. The variation of σ_w, normalised with the local U_*, with local stability, measured 7.8 m above the summit of the hill. The full line is the model prediction after Berkowicz and Prahm (1984), and the dashed line is the free convection prediction.

versus the local dimensionless stability parameter $-z/L$, where L is the local Monin–Obukhov length. The increasing trend at larger instabilities is consistent with the behaviour observed for flat, homogeneous terrain. The dashed line is the free-convection prediction whereas the solid line represents the recent model prediction of Berkowicz and Prahm (1984). This model represents the velocity fluctuations in an unstable boundary layer with orthogonal trigonometric functions and describes the buoyancy and shear-generated turbulence field with the relation

$$(\sigma_w^2/U_*)^2 = (b_w/K^{2/3})(-z/L)^{2/3} + a_w\,, \tag{4}$$

where σ_w is the standard deviation of the vertical velocity component and a and b are constants. Comparison with flat terrain results yielded values for a_w and b_w of 1.20 and 1.54, respectively. This implies a neutral limit for $\delta_w/U_* \simeq 1.10$. This value is somewhat smaller than normally found for non-uniform terrain; for example, Beljaars *et al.* (1983) obtained a value of 1.26. The neutral limit for the Great Dun Fell data indicates a value of 1.32. The Berkowicz and Prahm model also predicts the behaviour of the horizontal components, with a_w and b_w replaced in Equation (4) by $a_{w,v} = 4.0$ and $b_{u,v} = 0.3$, i.e., a neutral limit of 2.0, for each component. This compares with 2.05 and 1.78 observed

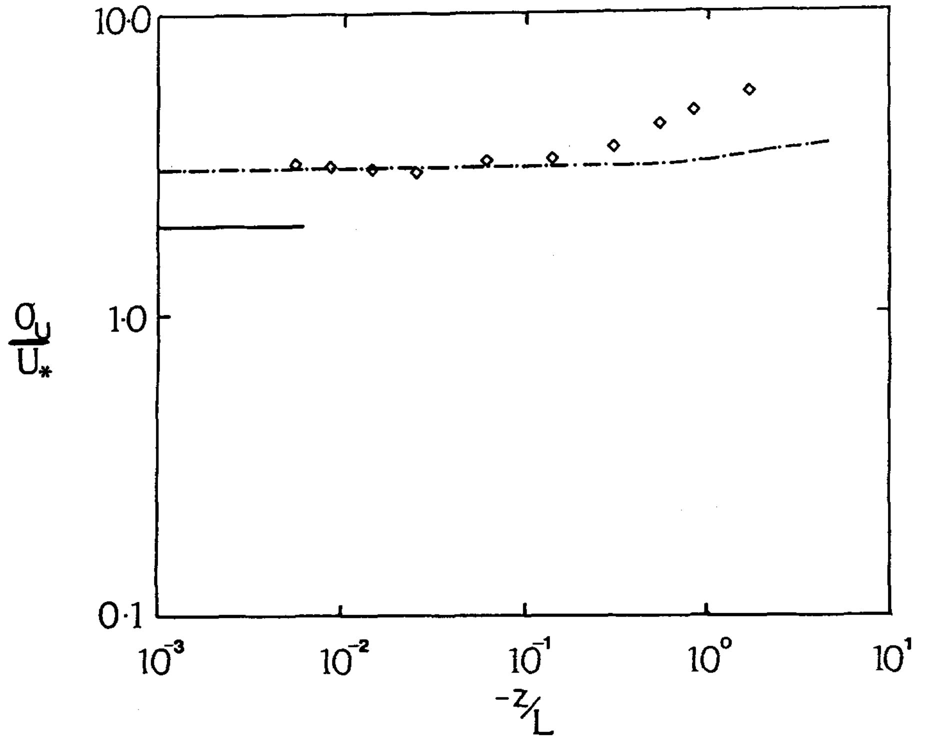

Fig. 8. The variation of σ_w, normalised with the local U_*, as a function of the local stability, measured at 7.8 m above the hill summit. The dashed line is the Berkowicz and Prahm model with a modified neutral intercept. The full line shows the neutral flat terrain intercept.

by Beljaars *et al.* for u and v, respectively. Figure 8 shows the stability band-averaged data for the u-component for Great Dun Fell. The dashed line is the model prediction of Berkowicz and Prahm adjusted to fit the neutral limit of this data set, which was 3.1 ± 0.4. The departure from the model at lower stabilities is probably due to the use of the local scaling parameter z/L rather than the more appropriate h/L, where h is the thickness of the convective boundary layer; see, e.g., Panofsky *et al.* (1977). Similar behaviour was observed for the transverse component, which gave a neutral intercept for $\sigma_v/U_* \simeq 2.78$.

All the above values are within the scatter generally observed over complex terrain, and in this respect the GDF data appear to be similar to those found over smaller hills.

6. Spectral Behaviour

6.1. General

Time series of the u, v, and w components were analysed in a manner similar to that described by, for example, Kaimal *et al.* (1972). The data were sampled at 40 Hz and block-averaged to 20 Hz to remove aliasing effects. Smoothing of the subsequent spectra was performed by averaging successive spectra. This generally produces good results and is preferred over smoothing filters which produce unwanted roll-off at lower frequencies, and may mask individual features of interest. In general, comparison with model spectra was done by smoothing the curves by eye. No detrending of the data was performed since low-frequency behaviour was of interest here. Kaimal *et al.* (1982) discuss the effects of long-term trends on such data sets in detail, large swings in behaviour occurring at the very lowest frequencies in general due to large changes in wind direction. The spectra presented here are confined to generally stationary conditions with respect to wind speed and direction and any contributions from trends should be small. High-frequency spectral loss was corrected by using approximate functions; see, for example, Kaimal *et al.* (1969) and Moore (1986). These simply restored the '$-\frac{5}{3}$' slope of the spectra in the inertial subrange.

All measurements were made at a height of 7.8 m, which is above any small effects that might occur due to slope change induced internal boundary layers at the front lip of the summit edge. Periods where wash effects from nearby structures occurred were generally very obvious in terms of high-turbulence levels, large variations in the local stress, and significant peaks in the computed spectra on scales of the order 3–5 m. These periods were rejected from the analysis. As discussed above, some measurements were also made at other sites on the hill remote from the buildings (40 m away). These data showed the same features to those presented below, and as a consequence we are confident that these data have not been contaminated by the presence of the buildings.

The velocity spectra were scaled using the Panofsky criterion, i.e., $(-u'w'U_*)^{2/3}$. As discussed previously, U_* was obtained from 5 and 10 m cup anemometers which compared well with the local stress as obtained from the sonic anemometer. This scaling criterion has been established for both Black Mountain (Panofsky *et al.*, 1982) and Maglarp Hill (Smedman and Bergstrom, 1984) as well as for transitions from water to very rough sloping land transitions (Högström and Smedman–Högström, 1984). Normalising the GDF spectra using this parameter produced inertial subrange levels to within less than 30% of the expected Kaimal model flat-terrain value. The high-frequency end of the spectra was somewhat noisier than found over much more ideal terrain but not to the extent where estimation of the inertial subrange proved difficult. As already mentioned, profile measurements have confirmed that local equilibrium is closely maintained through the lowest 3 m. Comparison wxith the anemometer measurements made at 5 and 10 m confirms that this equilibrium is maintained up to at least a height of 10 m.

For convenience in comparing with model spectra, the GDF spectra have been normalised here by taking $nS(n)_u/(U_*^2\phi_\varepsilon^{2/3}) = 0.3$ for the u-spectrum and 0.34 for the w- and v-spectra, where ϕ_ε is the dimensionless dissipation rate. The value of U_* estimated from this procedure was within the error bound mentioned above.

6.2. Unstable Stratification

Figures 9 shows spectra obtained from two periods when the wind was from the SW, i.e., perpendicular to the ridge line. Conditions were very unstable with low wind speeds and a mean local stability parameter $-z/L$ of approximately 0.35. A large excess in low-frequency energy occurs particularly in the horizontal components. The inflexion observed in the transverse spectrum is very characteristic of this component (Kaimal,

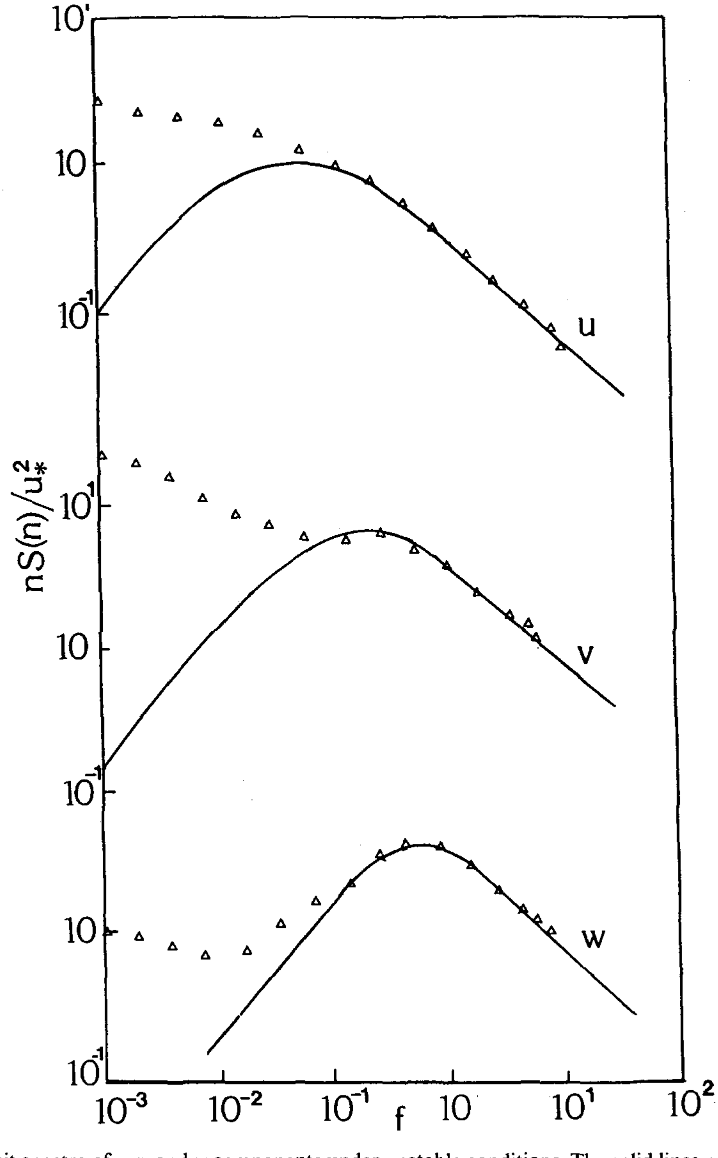

Fig. 9. Summit spectra of u, v, and w components under unstable conditions. The solid lines represent the Kaimal (1972) neutral spectral model for flat terrain. Each open symbol represents a band-averaged value of logarithmic spectral energy, normalised with the local U_*.

1976), which exhibits properties of both the longitudinal and vertical turbulence spectra. The behaviour is very similar to that observed over flat homogeneous terrain. This is in accordance with most measurements in complex terrain for unstable conditions, since terrain features become relatively unimportant in determining the gross spectral shape.

6.3. Neutral-Stable Conditions

Figures 10a–c show the u, v, and w spectra obtained under slightly stable conditions, $z/L \simeq 0.05$, as indicated by the low-frequency end of the w-spectrum which falls below the Kaimal model curve. The spectral energy recovers again somewhat for this component beyond $f < 0.01$, a region where gravity waves start to become important. The u-component shows a deficit for scales greater than 30 m. This feature was usually found to be well defined, with a rapid roll-off in spectral energy before it recovers again. This so-called 'deficit' is due to a combination of flow over a hill with the enhanced small scales, in local equilibrium, and damping of intermediate spectral scales by upper-level stable stratification.

6.4. Discussion of Turbulence Spectra

The low-frequency departure from the expected flat terrain spectral behaviour, as predicted by JH theory, was shown to occur for Black Mountain and also to some extent for the White Sands Ridge (Panofsky *et al.*, 1982). However, the Black Mountain spectra are very much smoother in their transition from inner-layer to outer-layer spectral scales with no recovery at lower frequencies. In addition, the terminology referring to 'deficit' is misleading, especially when no upwind spectra are available for comparison. A more correct term is 'spectral lag' inherent in which is the differing response times of the inner and outer layers. This was first introduced by Högström *et al.* (1982) in response to the prevalent 'large scales have longer memories' idea. Also inherent in the term 'spectral lag' is the idea that given sufficient unchanged fetch conditions, even the longest scales will eventually reach equilibrium as was demonstrated by the longitudinal spectra observed by Bowen and Lindley (Panofsky *et al.*, 1982) over small-scale escarpments. Over a hill, however, the inner layer is maintained over the entire length of the hill (Hunt and Richards, 1984), so that the larger scales reach equilibrium compared to flat terrain levels.

The region where spectral lag occurs for this site is marked SL in Figure 10a. The region marked T is the region where local slope changes will increase turbulence levels. The region marked L is the spectral region where wave contributions are likely to occur.

The results from two further case studies are shown in Figure 11. Two sets of u and w spectra are shown. Both cases were for high wind speeds $\simeq 17$–20 m s^{-1}, again for SW winds. Local stability varied from near neutral to moderately stable ($z/L \simeq 0.2$). The lag effect is not as well defined as in the previous cases. A slight deficit is noticed in the slightly stable case with enhancement at low frequencies. As the local stability increased slightly, the deficit became more apparent with greater variation in energy near the demarcation point. Interestingly, the vertical spectrum also showed enhancement

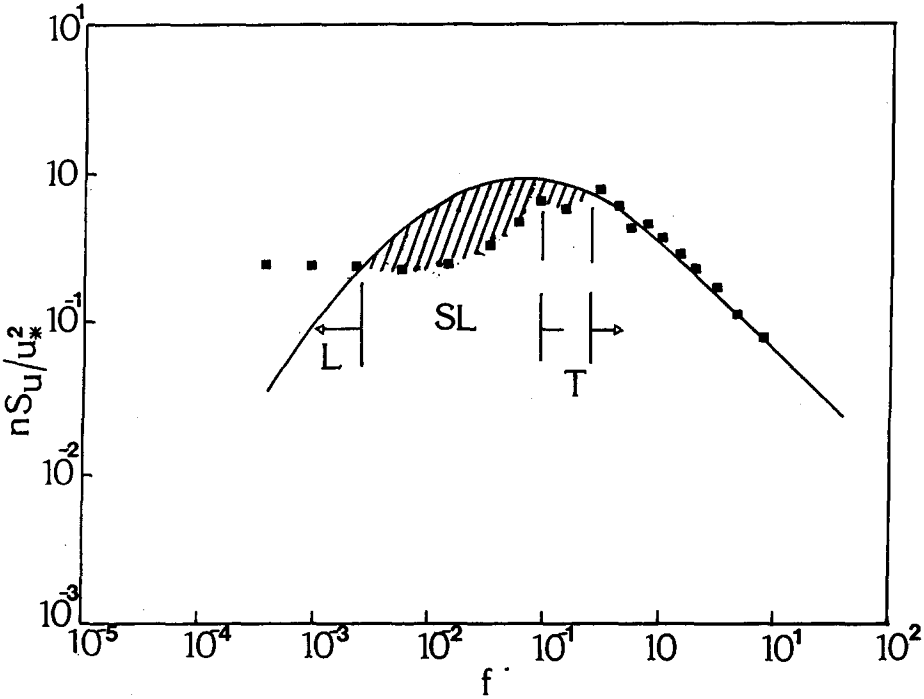

Fig. 10a.

Fig. 10a–c. Summit logarithmic spectra of *u*, *v*, and *w* components observed under neutrally stable conditions. The solid lines represent the Kaimal *et al.* (1972) model spectra. Figure 10a shows the various regions of interest; '*T*' is the region of terrain enhancement of turbulence, '*I*' is a possible transition region, '*SL*' is the spectral lag region, and '*L*' the region influenced by gravity waves.

at lower frequencies. These effects may be attributed to changes in upper-level stability which favoured the presence of lee waves.

One important difference between the spectra observed at this site and the theoretically predicted behaviour occurs in the case of the transverse component. This component also shows a spectral lag effect again for scales ≥ 30 m. This is contrary to what one expects from a knowledge of how the outer-layer turbulence field is deformed. The only other transverse spectra presented to date are again by Panofsky *et al.* (1982) for the White Sands Ridge, New Mexico. Although somewhat similar in appearance to GDF (for the SSW profile), it is concluded that the effective fetch is only 200 m, still long enough for distortion of large-scale eddies. No actual inner-layer depth is given but the spectra presented show very similar behaviour to that observed at Black Mountain. The transverse spectra, however, show no such distortion, being in close agreement with the Kaimal neutral model curve. Similar spectra for the E profile show no such distortion in either component. This strongly suggests that the spectral lag effect dominates over the rapid distortion effect at our site.

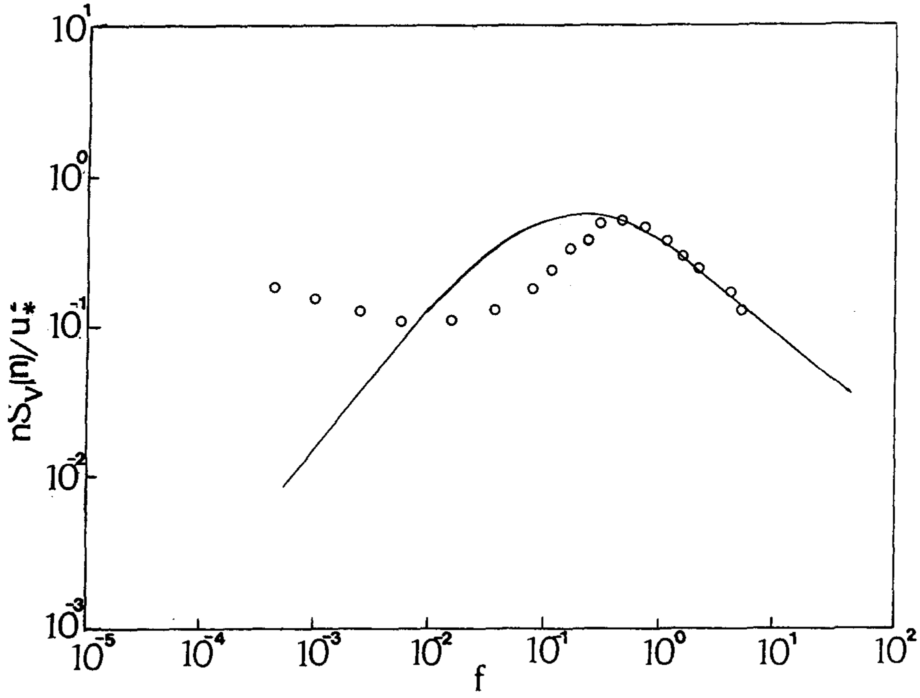

Fig. 10b.

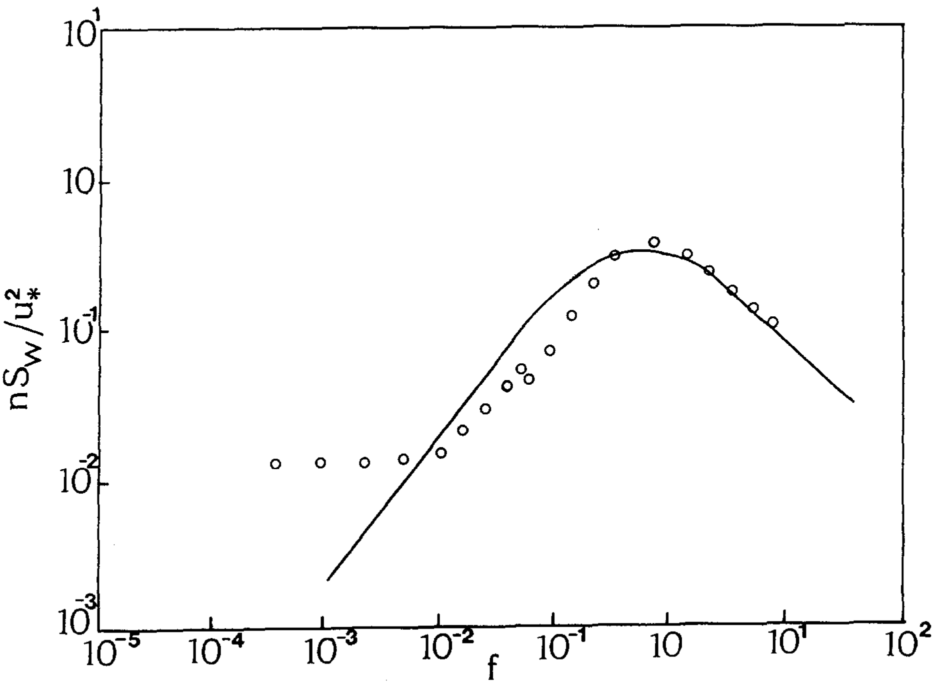

Fig. 10c.

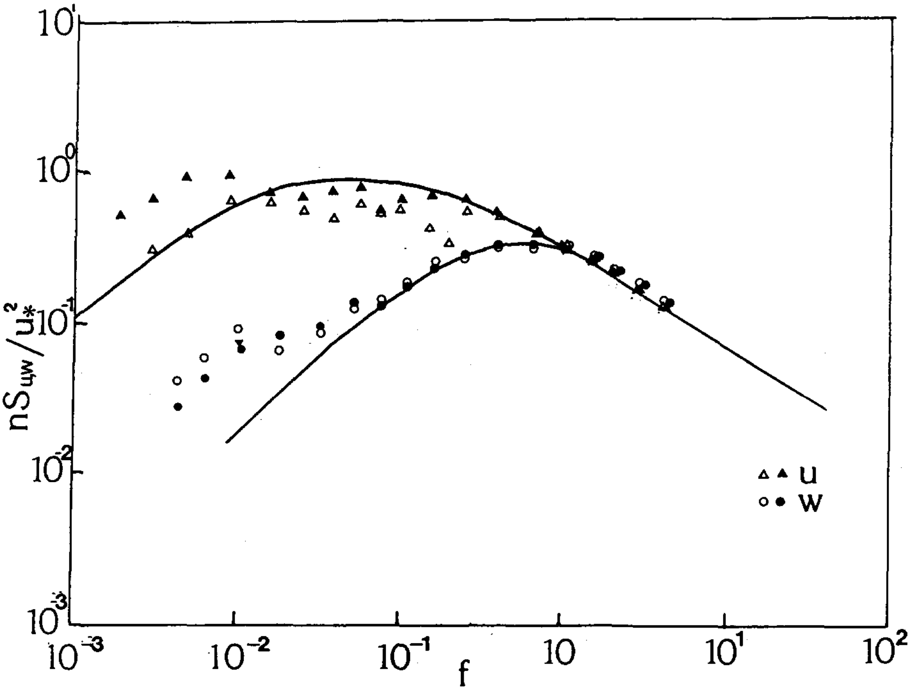

Fig. 11. Spectra of u and w observed at the summit of GDF under locally neutral conditions. The curves represent the Kaimal (1972) spectral models. The open symbols are for wind speeds $\simeq 17$ m s^{-1}, while the solid symbols represent spectra for wind speeds $\simeq 20$ m s^{-1}.

Similar results were found by Peterson *et al.* (Panofsky *et al.*, 1982) at Risø, downwind of a lake-sloping land transition. Also Beljaars *et al.* (1983) indicate that both horizontal components of turbulence are subject to slow relaxation and both respond to changes in slope and roughness in a similar manner.

If it can be assumed that the point of spectral departure from the Kaimal model does represent the inner layer of GDF, then there remains a discrepancy between the predicted value of $\simeq 80$ m, from JH, and the observed value of $\simeq 30$ m. Until recently, the very limited data tended to support the predicted inner-layer relation well. Black Mountain differed substantially from GDF in that it was covered with extensive forestation, requiring a large zero-plane displacement for the logarithmic profile ($\simeq 8$ m). Spectra from Maglarp hill although confirming two quite different inner-layer depths, showed no deficit or lag effects but rather a substantial enhancement at low frequencies. Bearing these difficulties in mind, it should be emphasized that the relationship arising from matching the inner and outer layer equations is approximate since advective terms are neglected; these may slow the response of the turbulent flow to changes in slope. In other words, the turbulence in the inner equilibrium layer may not be propagated as rapidly as predicted, resulting in a smaller value of l. Hunt and Richards (1984) recently indicated that the stress profile is much greater near the surface than throughout the bulk of the inner layer. The recent work of Mason (1986) confirms this. His work, which is the only one in which measurements have been made sufficiently close to the surface without interference from local terrain changes, suggests that l is substantially smaller than suggested by JH.

More recently, Zeman and Jensen (1987) have provided a more rigorous treatment of turbulent airflow over a hill using second-order closure equations. Their model predictions were confirmed by measurements over a small hill, Askervain, in the Outer Hebrides. This model predicts an inner layer depth $\frac{1}{3}$ that of the JH value, and the turbulence is only in equilibrium within the lower half of the layer. Using their equations, the inner-layer depth for GDF is about 27 m, which is consistent with the measurements presented here. They also show that the effects of curvature and rapid distortion combine to produce complicated profiles in the outer layer, particularly for the vertical component. In addition, the rapid distortion effect is very dependent on upstream anisotropy. The discrepancy between the Black Mountain and Maglarp hill results and the Zeman and Jensen predictions indicates some difficulty in observing inner layers over all but the simplest cases.

7. Conclusions

The following conclusions may be drawn from this study.

(1) The speed-up of the wind over Great Dun Fell is dominated by the elevated stratification but generally agrees closely with predictions of Jackson–Hunt theory extended to include elevated stratification by Carruthers and Choularton (1982). Notable exceptions, however, occur when the flow is blocked.

In contrast to a low hill, no clear relationship exists between speed-up and local stability measured near the surface at the summit site. This is because over a high hill, the speed-up is dominated by the structure of stratification in the troposphere.

(2) Within 10 m of the hill surface, the turbulence is in local equilibrium with the local terrain in neutrally stable conditions. The longitudinal and transverse components show spectral lags at wavelengths greater than about 30 m. This compares with a predicted inner-layer depth of 80 m from Jackson and Hunt (1975), 15 m from Mason (1986), and 27 m from Zeman and Jensen (1987).

(3) Turbulence components normalised with the local U_* are slightly larger than those found over flat terrain but compare well with most complex-terrain values. The behaviour at larger instabilities agrees well with the model of Berkowicz and Prahm (1984).

(4) At low-reduced frequencies, $f \leq 10^{-1}$, a recovery in spectral energy is apparent due to gravity wave activity. This may be linked to lee wave activity generated by the hill.

(5) Wind tilt at the summit was generally close to zero. However, values of up to $+6°$ occurred associated with flow separation to the lee of the hill while in strongly supercritical flow, a downward tilt was sometimes observed.

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