THE RELATIONSHIP OF WIND STRUCTURE TO WIND LOADING

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SUMMARY

TO evaluate wind loads on structures a knowledge of wind structure on both a macro- and micrometeorological scale is necessary. The former is concerned with the overall climatic properties of the wind and the recurrence of extremely destructive winds: the latter is concerned with the details of the flow in the earth's boundary layer - the wind velocity profile and the gustiness. Both aspects are discussed.

In a neutrally stable atmosphere (appropriate to strong winds) properties of the mean flow are shown to be dependent almost entirely (except near mountains) on the roughness of the ground which in nature can vary radically between open water and heavily built up urban areas. Simple power law velocity profiles with appropriate exponents and having a common gradient velocity at the upper limit are found to give good approximations to observed and predicted profiles over surfaces of widely differing roughness. The transition in boundary-layer flow which occurs when the wind blows from one surface to another of different roughness is examined.

The statistical properties of turbulence in strong winds are described in terms of the spectra and scales as well as the probability distributions. It is suggested that reasonably reliable estimates of these quantities can be made from a knowledge of the mean wind velocity and the ground roughness. A "two-dimensional jet" model of boundary-layer flow in the wind tunnel suggested by Townsend is shown to correspond to many of the properties of turbulence in strong winds.

The problems relating to the establishment of an "extreme-mean-wind-velocity" field over a country are discussed. For this purpose it is suggested that the use of a gradient-wind-velocity field has many advantages to offer. A map of the extreme-hourly-gradient-wind parameters over

the United Kingdom is shown as an illustration. From this, surface wind velocities (and hence the gust spectra) can be found if the ground roughness can be estimated. The advantages of site-tests for determining wind properties are pointed out.

1. INTRODUCTION

Although wind loads are a major factor in the design of many structures, the nature of the wind itself is not a subject with which structural engineers are generally familiar. This state of affairs is no doubt partly due to the interdisciplinary nature of the subject, and partly because of the lack of emphasis usually given to it in structural engineering curricula. As a result however, structural design has tended to become compartmentalized: the estimation of wind-loads is often divorced from the study of the structure itself and the responsibility for their estimation delegated to others. Indeed, to a few structural engineers destructive winds may be little more than unpredictable acts of fate capable of little or no scientific description!

This somewhat defeatist approach may explain why apparently so little incentive has been given to research in the meteorological areas related to wind loading while in more clearly defined areas - such as the estimation of pressure distribution on structures from wind-tunnel models - much more effort has been expended. Next to no research, for example, has been carried out into the wind structure over an urban area in spite of its obvious relevance. Only recently has the climatology of high winds been examined. Subjects such as the influence of terrain roughness and topography on the wind properties have been almost untouched (with the exception of one or two notable contributions such as that due to Jensen) and certainly have had no impact on wind loads.

The intention in this paper is to try to point out the possible direction a scientific formulation of the wind properties relevant to structural design might take. This cannot be done, however without first establishing those features of structural behaviour which are liable to have a bearing on the subject and which define the terms of reference.

1.1 Wind-load criteria

The most serious consequence of wind action, the collapse of the structure, can arise either through a single application of a stress greater than the structural material can sustain at a sufficient number of points to render the structure a mechanism, or it can arise through repeated applications of somewhat lower stress levels which gradually lead to the fatigue failure of the members. Both types of collapse have occurred: the Tay Bridge disaster of 1879 is only one of several examples amongst large bridges of the former type of failure due to wind: examples of fatigue failure have been mainly amongst flexible structures such as street lamp standards and the hangers of suspended bridges as in the instance of an arch bridge at Tacony-Palmyra in 1929.

There are other consequences of wind loading which also must be considered, such as large deflections or accelerations. Although not a direct criterion of structural failure these can often sufficiently impair the utility of a structure, to be a primary design consideration. One can find examples of this in the excessive deflection of directional radio antennae, the clashing of power line conductors, and in the feelings of insecurity that large deflections and accelerations produce on occupants of tall buildings and bridges.

All these effects of the wind have this much in common; they all depend not only on the magnitude of the load but also on the time sequence of its application. This last fact which is a consequence of the dynamic characteristics of structures, is not widely recognized by structural engineers (or meteorologists) although its significance is greatly increasing as discussed in another paper.

A further factor which must be considered and which is directly related to the economic utility of a structure is its anticipated useful lifetime. It is seldom that this can be stated explicitly (except in the rather artificial sense used by accountants in writing off a structure in their books) although it is very much a fact of life that some structures are intended to last much longer than others. A cathedral is expected to last longer than a stop-gap "prefab" house or a radio mast in which electronic obsolescence is likely to have set in within 20 or 30 years. It is clearly economically desirable to reflect these differences explicitly in structural design and correspondingly in the wind loading.

These factors define the terms of reference in formulating design wind loads. It emerges that what is required is a statement defining first the magnitude of the wind speed at the site of the structure, second its time dependence both on a time scale characteristic of its lifetime and also of its dynamic response, and third its spatial characteristics in the region of aerodynamic influence surrounding the structure.

2. THE STRUCTURE OF THE WIND

2.1 General

The source of the wind's energy can of course be traced back to the sun. In very simple terms solar radiation, which is far more intense at the equator than the poles, tends to cause differential heating of the earth's surface which in turn gives rise to gradients of pressure in the atmosphere. These gradients of pressure can be inferred from the contours of equal barometric pressure (isobars) shown on weather maps such as that shown in Fig.1. At heights greater than 1000 ft or so, outside the influence of the frictional forces near the ground, the wind attains what is known as the gradient velocity. Its magnitude is a function of the latitude, the curvature of the isobars and their spacing i.e. pressure gradient.

Closer to the ground the airflow is slowed down by the drag forces at the surface and which are transmitted upwards by the virtual Reynolds stresses which result from the momentum exchange between layers due to turbulence. The rougher the ground the greater the drag force, the Reynolds stresses, the turbulent intensity, the retardation at the surface and the gradient height.

It follows that the wind at a point near the earth's surface will be characterized first by the large-scale movements of the pressure systems giving rise to the gradient wind and second by the modifying influence of the ground surface. It turns out that the two processes give rise to variations in wind speed having completely different time scales. This is indicated in Fig. 2 in which the spectrum of horizontal wind speed near the ground is shown for an extended range of frequencies.

Such a spectrum is analogous to the spectrum of light formed by passing light through a prism and which expresses the distribution of light energy with wavelength. One method of obtaining a wind-speed spectrum is to convert the wind-speed signal into its electrical analogue pass this through a range of filters having different characteristic frequencies and measure the output energy on a watt meter. In Fig. 2 the spectrum is plotted on a logarithmic frequency scale in such a way that the area under the curve between any two frequencies is proportional to the energy within that frequency range.

The spectrum is seen to be characterized by several prominent peaks, with several intervening ranges of low amplitude. On the right, the high frequency end, there is a broad peak with a maximum at a period of between 1 and 2 minutes per cycle. This part of the curve (which was in fact evaluated from records made in hurricane Carol) is characteristic of turbulence generated almost entirely mechanically by shear stresses at the ground surface: in high winds convection plays a minor role not only because the air is so churned up that thermal instabilities do not get a chance to build up; but also because the shear stresses and mechanical turbulence build up in proportion to the square of the wind speed and very soon swamp any contribution that convection might otherwise make even if conditions were favourable.

Between periods of 5 minutes and about 4 - 5 hours the spectrum contains very little energy implying that there is very little variation in wind speed occurring at these rates. At a twelve hour period there is a minor peak corresponding to the lulls in wind speed which generally occur at sunrise and sunset. At a period of about four days there is a major peak corresponding to the movement of large-scale weather systems such as those in Fig. 1: everyday experience tends to confirm that weather conditions fluctuate with a period of this order.

Spectral estimates for periods greater than about two months were not made by Van der Hoven but experience indicates that another very predominant peak occurs at a period of a year. This corresponds to seasonal fluctuations in wind speed which come about from the greater average difference in temperature between the equator and poles (which is the effective operating range of the earth acting as a heat engine) during winter than in summer. Other peaks might in fact be present for diurnal and lunar cycles.

Beyond a period of one year the shape of the spectrum is speculative. However it is difficult to conceive any other physical process capable of causing longer period fluctuations other than the eleven year sun spot cycle which can only be a weak effect.

The concept of atmospheric motion as compounded of superimposed trains of wind waves of different frequencies and of amplitudes dictated by the spectrum is a useful one to have. An extremely strong wind can then be visualized as the simultaneous arrival of the peaks of several important wave trains. If the probability distribution, defining the proportion of the total time the wind speed exceeds a given value, were known - as well as the spectrum - it would theoretically be possible to estimate such quantities as the time interval between recurrences of extremely high wind speeds, which in part at least is the object of our inquiry. This approach does not however turn out to be very fruitful mainly because of the difficulty of estimating the left hand, low frequency portion of the spectrum; also the same results can be derived more directly as will be explained. The immediate usefulness of this approach lies in its aid to understanding of the processes involved.

It is important to realize that although a spectrum such as Fig. 2 will only apply to a particular site and to a particular height above ground, the general form of the spectrum and the position of the peaks may nevertheless be expected to remain very much the same regardless of the geographical locality, the nature of the terrain and the height above ground. One of the most important distinctions that it appears can be made is between the fluctuations of a macrometeorological kind such as the movement of large-scale pressure systems, seasonal variations etc. and those which are of a local, micrometeorological kind and associated with the flow characteristics of the boundary layer itself. In more simple parlance these two categories of fluctuation might be described as "weather-map fluctuations" and "gusts."

From Fig. 2 it appears that these two types of fluctuations are separated by a gap extending from roughly five minutes to five hours. This gap is important to our evaluation of wind loads for several reasons. It enables a clear cut distinction to be made between gusts and weather-map disturbances and furthermore their causes: in another sense this distinction can be regarded as between gusts and the mean wind where the mean wind is characterized by the average velocity over some period within the spectral gap. Fortunately almost all routine meteorological measurements of wind speed - with the exception of so-called peak gusts - are averaged over periods within the spectural gap. Mean hourly wind speeds are for example recorded in the U.K. and Canada, mean five-minute speeds in the U.S.A., mean ten-minute speeds in Japan etc. Because of the low amplitudes in the spectral gap all these various average speeds are in high winds almost equal. Thus it would seem that the meteorological measurements made over many years have been referred to averaging periods which (by good luck or good judgment) happen to be most suitable for making long term estimates of wind loads. From the experimental point of view the spectral gap enables stable estimates of the gust spectrum to be made from records of fairly short duration - 20 minutes or so. Because structures have natural periods of vibration of less than 1 minute it turns out that the part of the spectrum relating to gusts has high importance in the evaluation of wind loading.

The existence of this gap although not proved incontrovertibly, now seems strongly probable, certainly in high winds when, as explained earlier, lower frequency convective turbulence is relatively speaking far less significant. Van der Hoven demonstrated that it existed in a wide variety of weather conditions and localities and the writer has confirmed its existence in a run of very high wind measured at the Severn River railway bridge in Gloucestershire (kindly loaned by Messrs. Freeman, Fox and Partners.[11]) It is attributed to the lack of any physical processes capable of generating fluctuations at gap frequencies.

If under some circumstances the gap does not exist and the "weather map," convective and other non-mechanical disturbances extend the macro-meteorological range of the spectrum so that it overlaps and augments the gust spectrum, the task of evaluating the dynamic gust response will become problematic and the long-time meteorological records of hourly or five-minute-average winds of far less value. Fortunately however, this does not appear to be true, and, in the opinion of the writer, the model upon which wind loads are based should at present, assume the existence of this gap in the overwhelming majority of destructive winds.

The conditions under which this assumption might be in greatest error are intense local storms such as severe frontal squalls, thunder-storms and, in the extreme case, tornadoes. Although these types of storms are known to occur frequently in several regions of the world they are generally of such local extent, affecting only one or two square miles at a time, that the likelihood of a destructive storm of this type striking a particular site is negligibly small. This is evinced partly by experience in the notorious tornado belt of the U.S. midwest where it is never considered necessary to design against their contingency. It will be assumed in this discussion, that local storms of these types are either sufficiently rare or otherwise harmless that, ignoring the special design conditions they may present, will not affect the overall statistical picture of the wind loading arrived at from a study of large scale destructive storms for which the assumption of a spectral gap seems valid.

The model of the wind which emerges from this is one in which the overall flow characteristics (indicated by the gradient wind) are governed by large scale, slowly varying macrometeorological processes having little significant periodicity less than two or three hours. Nearer the surface this flow is modified by the boundary layer in which the mean flow is retarded and mechanical turbulence is generated.

Our model of the wind near the ground is thus analogous to that near the boundary layer of a wind tunnel in which the speed of the tunnel fan is slowly varied, at a rate which, (by analogy to the spectral gap) is much slower and of longer periodicity than any turbulent fluctuation. To take the analogy further the roughness of the wind-tunnel wall should be variable to correspond to the full-scale roughnesses of open water at one end of the scale and a built-up urban area on the other end. Under these conditions the mean velocity at a given height in the boundary layer will be proportional to the velocity at the tunnel centre line and will vary at the same rate, its magnitude will be a function of the roughness of the surface and hence the boundary-layer profile. The mean square of the fluctuation velocity in the boundary layer will depend on the surface shear stress which in turn depends on the surface roughness and the square of the mean velocity. The same general characteristics may be expected to pertain in high winds.

3. CHARACTERISTICS OF THE MEAN FLOW IN THE EARTH'S BOUNDARY LAYER IN HIGH WINDS

3.1 General

Above the layer of frictional influence near the surface the air moves purely under the influence of the pressure gradients and attains what is known as the gradient velocity. From the equation of motion for the air it is found[30] that the magnitude of the gradient wind round a low-pressure region (inducive of higher winds) is given by

$$V_G = R\,\omega \sin\lambda \left[\sqrt{\frac{\frac{dp}{dn}}{\rho R\,\omega^2 \sin^2\lambda} + 1} - 1 \right] \qquad (1)$$

in which $\frac{dp}{dn}$ is the pressure gradient, ω is the angular velocity of the earth, λ is the latitude, ρ is the air density, and R is the radius of curvature of the isobars. The direction of the wind outside the region of frictional influence is parallel to the isobars as shown by the arrows in Fig. 1. Usually the fraction inside the square root sign is considerably less than unity and by binomial expansion we can write

$$V_G \approx \frac{\frac{dp}{dn}}{2\rho\,\omega\sin\lambda}\left\{1 - \frac{\left[\frac{\frac{dp}{dn}}{2\rho\,\omega\sin\lambda}\right]}{2\,R\,\omega\sin\lambda} + \right. \qquad (2)$$

The second term in the expansion is only significant for small radii of curvature, and under most circumstances a good estimate is given by the first term

$$V_G \approx \frac{\frac{dp}{dn}}{2\rho\,\omega\sin\lambda} = \frac{\frac{dp}{dn}}{\rho f} \qquad (3)$$

where $f = 2\,\omega\sin\lambda = 1.458 \times 10^{-4}\sin\lambda \ \ sec^{-1}$.

This is known as the geostrophic wind and is that associated with the Coriolis acceleration due to the rotation of the earth.

The height at which the gradient velocity is attained will be denoted by Z_G and is generally of the order of 1000-2000 ft.

Closer to the ground the airflow is slowed down by the drag forces at the surface and the virtual Reynolds stresses which result from the momentum exchange between layers due to turbulence. In high winds the boundary layer is fully turbulent over all natural surfaces; the Reynolds stresses exceed the direct viscous stresses by several orders of magnitude and the latter may be neglected. Viscosity does, however, play an important role in controlling the rate of dissipation of the turbulence and hence indirectly the Reynolds stresses also. The rougher the ground, the greater the drag force at the surface, the turbulent intensity, the Reynolds stresses, the gradient height and the retardation at the surface.

The overall region of frictional influence below Z_G is termed the planetary boundary layer. It can be broken down into at least two sub-layers, principally the surface boundary layer, in which the shearing stress is approximately constant, and a transition region in which the shearing stress falls off from the constant value of the surface layer to the practically zero value in the free atmosphere.

In the surface layer which extends up to roughly 200 ft (± 100 ft) the wind velocity profile appears to be accurately defined by the Prandtl logarithmic profile. This is based on the assumption first of constant shearing stress and second of a so-called mixing length (analogous to the molecular mean free path) proportional to the height. It is valid in conditions of neutral stability (already pointed out to be appropriate for high winds) and is written

$$\frac{V_Z}{V_*} = \frac{1}{k} \ln \frac{Z}{Z_o} \tag{4}$$

where V_Z is the mean velocity at height Z, $V_* = \sqrt{\frac{\tau_o}{\rho}}$ is the "friction velocity", Z_o the roughness length, k von Karman's constant (≈ 0.4), τ_o is the shear stress at the surface and ρ the air density. Related to the friction velocity is the surface drag coefficient κ such that

$$\tau_o = \kappa \rho \bar{V}^2 \tag{5}$$

and

$$\kappa = \left(\frac{V_*}{\bar{V}}\right)^2 = \frac{k^2}{\ln^2 \frac{Z}{Z_o}} \tag{6}$$

The logarithmic profile agrees well with experimental measurements in the wind-tunnel boundary layer and in the surface layer over natural surfaces of roughnesses varying between smooth mud flats, open water and thick grass. It has yet to be demonstrated experimentally to be characteristic of urban areas. Above the surface layer the effect of the Coriolis force increases, the effect of the surface roughness decreases and the profile departs significantly from the logarithmic form.

Several approaches have been taken to formulate the mean-wind profile through the deeper planetary boundary layer up to the height at which there is no systematic deviation from the wind motion implied by the barometric pressure gradient and geostrophic forces. Most of these have introduced the concept of the eddy viscosity. This is a means for defining the virtual shear stresses which arise from the exchange of momentum between layers. Precise formulations have generally floundered on the problem of defining the variation of the eddy viscosity with height. An assumption of constant eddy viscosity has led to the so-called Ekman spiral. Other models assuming power-law variations of the eddy viscosity with height have been put forward by Prandtl and Tollmien and Kohler[30]. A two-layer model which implies a mixing length which increases linearly with height to the top of the surface layer and above that decreases linearly to the top of the planetary layer, has been suggested by Rossby and Montgomery.[26]

These solutions account for several observed characteristics such as the systematic deviation of the wind direction from that of the isobars nearer the ground (greater over rougher surfaces). In spite of their greater sophistication, however, these expressions all contain quantities which can at present only be defined empirically. Their overall reliability in providing numerical predictions of the wind-speed profile do not appear to be greater than the simple power law profile given by

$$\left(\frac{\bar{V}_Z}{\bar{V}_1}\right) = \left(\frac{Z}{Z_1}\right)^{\alpha} \tag{7}$$

in which $\bar{V}_1$ is a reference velocity at height Z_1 and α is a constant. Because of its simplicity and the lack of any expression yielding better accuracy, this seems a suitable profile for wind loading purposes.

If this curve is to represent the velocity throughout the boundary layer then at some height Z_G the velocity must attain the gradient value V_G. In this case

$$\frac{\bar{V}_Z}{\bar{V}_G} = \left(\frac{Z}{Z_G}\right)^{\alpha} \tag{8}$$

From this, if the index α and Z_G are known the wind velocity at any height in the boundary layer can be expressed as a ratio to the gradient wind. Both parameters, it seems, are principally functions of the roughness of the ground.

To determine values of α and Z_G for different natural surfaces the writer in an earlier paper[9] has surveyed the published measurements of wind profiles in high winds for some 19 different localities ranging in roughness from open water to the centre of a large city. The measurements varied in quality, some being based on a series of measurements and made over a considerable height range, in other instances they were based on isolated measurements or only over relatively shallow height ranges up to 100 ft. From these data and other information relating to the ratio of gradient to surface winds emerged a reasonably consistent picture from which the average values of α and Z_G given in Fig. 3 were derived. The profiles typifying three types of terrain - open terrain, wooded country and an urban centre - are shown in Figs. 4 and 5.

The main body of evidence for these curves was given in the previous paper and it is not the intention to repeat this here: however, since this was written other information has come to light which relates to these curves.

A study by Johnson[22] of the relation between the mean (hourly) wind velocity at 2 metres, 40 feet and 1000 feet at the Suffield Experiment Station near Medicine Hat, Alberta, has shown that $\frac{V_{40}}{V_{1000}} = 0.60$ and $\frac{V_{2m}}{V_{1000}} = 0.32$ in conditions of neutral stability. The site is in the heart of the so-called "short-grass country" in the unrelieved flatness of the Canadian prairies. If the 1000 foot velocity is taken as representative of the gradient wind these ratios provide confirmation of the "flat, open country" curve of Fig. 5. This curve also agrees well with Taylor's measurements in strong wind over Salisbury Plain (quoted by Sutton in "Micrometeorology") from which the same ratio was found to be 0.61.

A revealing study of the action of surface roughness on the wind was undertaken by Jensen in Denmark and given in his notable thesis on "Shelter Effect" (Jensen - 1954). Anemometers were set up at 2m height along two east-west lines running from coast to coast across the Danish peninsula (see Fig. 6). The length of the lines were 47½ miles (76 km) and 63 miles (101 km) and instruments were spaced at roughly 4 and 6 mile intervals on the two lines respectively. Simultaneous measurements of wind speed were made at heights of 2 metres on each of the lines (2 sets on line 1 and 3 in the case of line 2) on the occasions when the gradient wind was from the west at approximately 20 m/s (45 m.p.h.).

In Fig. 6 the velocities at each of the anemometers are given as a ratio to the gradient wind. Also shown is what Jensen terms the "roughness of great order" along the lines. This was calculated from the average frontal area of the obstructions per unit horizontal area of terrain multiplied by an "effect figure" which ranged between 2 and 4 for coniferous hedges and trees and between 1 and 3 for deciduous (according to spacing) and 4 for buildings, gardens and small plantations. These "effect figures" were determined from wind-tunnel measurements of the wind profile downstream from series of screens of varying porosity.

These curves in Fig. 6 indicate clearly the strong influence of surface roughness on the wind strength and tend to confirm the curves in Figs. 4 and 5.

The effect of surface roughness on two further parameters, the roughness length Z_0 and the surface drag coefficient is also important (see Eqs. 4 and 6).

Values for Z_0 are usually estimated directly from the logarithmic wind profile near the ground. Representative values for roughnesses up to thick grass are quoted by Sutton in "Micrometeorology": for rougher surfaces we have to turn elsewhere. Measurements over wooded areas have been carried out at Brookhaven Laboratory and at Munich. At the former the accepted value for Z_0 is 1 metre[14] - while Lettau[21] quotes values of 20 cm from the profiles obtained by Baumgartner[2]. An analysis by Davenport[10] of some somewhat sparse data obtained by Jensen, and Shiotani over urban areas indicate that here the roughness length is of the order 1-3 metres.

Values of the drag coefficient have been measured in four ways:- from observations on the approach to the geostrophic wind, from wind profiles near the ground, by direct measurement using drag plates, by measurement of the Reynold's stress near the ground (mean product of longitudinal and vertical fluctuation velocities). For open sea Sutton[30] quotes Sutcliffe's value of $\kappa = 0.0005$ (with 10 metre reference height): for rough grassland 0.005 seems an accepted value: for a wooded area such as Brookhaven the roughness length 1m indicates $\kappa \approx 0.030$ while the Reynolds stress indicates $\kappa = 0.015$, which agrees with the value obtained from the wind profile for the Munich forest data mentioned above. If a roughness length of 3m is accepted for a city then the coefficient of drag turns out to be about 0.050.

These surface drag coefficients are, in fact, close to those encountered in artificial surfaces such as pipes, aircraft wings, and ships plating. It is also practice to represent the boundary-layer velocity profile over these surfaces by a power law (such as the well known 1/7 power law). Nunner[23] has shown (see Fig. 7) from pipe flow experiments that there is a systematic relationship between the index of the power law and the surface drag co-efficient, as defined by $\left(\frac{V_*}{V_{AV}}\right)^2$ (see equation (2)), which can be expressed empirically by

$$\alpha = \sqrt{8\left(\frac{V_*}{V_{AV}}\right)^2}$$

where V_{AV} denotes the average velocity in the pipe. Turning to the profiles derived from the natural wind - the parameters of which are given in Fig. 3 - it can be shown that the average velocity in the boundary up to the gradient height can be expressed in terms of the 10 metre reference velocity as

$$\frac{V_{AV}}{V_{10}} = \frac{1}{1+\alpha}\left(\frac{Z_G}{10}\right)^{\alpha} = \emptyset \quad \text{(say)}$$

and that

$$\left(\frac{V_*}{V_{AV}}\right)^2 = \frac{\kappa}{\emptyset^2}$$

Three typical values of this function are also plotted in Fig. 7 as a function of the corresponding power law exponent and are seen to be well within the scatter of the experimental results. This interesting result indicates the universality of the turbulent boundary layer properties on two vastly different scales and for two different media.

The main objections to the above power law approach to defining the mean wind velocity profile are its empiricism and its dependence on the non-fundamental parameters α and Z_G both of which have a somewhat nebulous physical reality. The justifications for adopting the approach are first simplicity and second the fact that both α and Z_G or functions of these quantities can, it seems, be systematically related to the fundamental parameters defining the roughness of a surface or otherwise its effects, principally Z_o and V_*. A more direct approach is to group together the various parameters which do have some physical reality using dimensional arguments: two parameters of significance emerge

$$\frac{V_G}{fZ_o} \quad \text{and} \quad \frac{V_*}{V_G} \quad \left(\text{or} \quad \frac{V_*}{fZ_o}\right)$$

All of these quantities are directly measurable (in contrast to α and Z_G which can only be inferred indirectly). V_G, the geostrophic wind, is directly translatable into the pressure gradient by the transformation

$$V_G = \frac{\frac{dp}{dn}}{\rho f} \text{ (see equation 3)} \tag{9}$$

where $f = 2\omega \sin \lambda$ (the ingredient of the Coriolis acceleration due to the earth's rotation): V_* can be directly translated into the shear stress at the surface: Z_o is the roughness length and constitutes a statistical measure of the average size, shape and arrangement of the surface obstructions.

The first of these parameters $\frac{V_G}{f Z_o}$ has been termed the Rossby number Ro. The second $\frac{V_*}{V_G}$ is by nature of a drag coefficient (or the square root of the more conventional form of drag coefficient such as κ) and Lettau[21] has termed it the geostrophic drag coefficient C_a. (The subscript denotes an adiabatic lapse rate i.e. neutral stability.)

As outlined, a number of theoretical models of the wind in the boundary layer have been suggested[26,30]. All are based on inherent assumptions (concerning the variation of eddy viscosity, mixing length etc. with height) which can explicitly or implicitly be written in the form

$$\frac{V_G}{f Z_o} = Ro = F_1 \left(\frac{V_*}{f Z_o}\right)$$

or alternatively

$$C_a = \frac{V_*}{V_G} = F_2 \text{ (Ro)} \tag{10}$$

Taylor[32] has shown that, in spite of the range of inherent assumptions, several of the theoretical solutions can be matched extremely closely by simple power law relationships of the type

$$\frac{V_G}{f Z_o} \propto \left(\frac{V_*}{fZ_o}\right)^m$$

which is directly equivalent to

$$C_a \propto Ro^p \text{ where } p = \frac{1 - m}{m} \tag{11}$$

(Taylor ascribes this similarity to the dominance of the boundary conditions on the result viz. - constant shear stress near the surface and zero shear stress as the gradient velocity is approached). The same boundary conditions are assumed in all models.

In the several practical instances and theoretical models Taylor found p to be approximately -0.09. The validity of this expression tends also to be confirmed by some data collected by Lettau[21] for localities of varying latitude and roughnesses ranging between open sea and forest. These are shown on a log. plot of C_a versus Ro in Fig. 8 and compared with the line for p = -0.09. The full expression for the line is

$$C_a = 0.16\ Ro^{-0.09} \tag{12}$$

To find the velocity of the wind at a height Z near the surface in terms of the gradient wind, it is now necessary only to multiply both sides of the logarithmic profile (equation 4) by $\frac{V_*}{V_G}$ yielding

$$\frac{V_Z}{V_G} = \frac{C_a}{k} \ln \frac{Z}{Z_o} \tag{13}$$

This expression is evaluated in Fig. 9 for heights of 10 m and 2 m as a function of the roughness length Z_o for the practical range of wind speeds (V_G = 25-50 m/s i.e. 55-110 m.p.h.) and for latitudes 30° and 60°. It is seen that the ratio is relatively insensitive to latitude or gradient wind velocity but highly sensitive to the roughness of the surface. Furthermore, the ratios of wind speed at the two reference heights agree quite well with the ratios suggested in the simple power law profiles of Figs. 4 and 5. The great advantage of the presentation of Eq. 13 and Fig. 9 is that due to the insensitivity to the latitude and wind speed the ratio $\frac{V_Z}{V_G}$ is determined almost uniquely by the roughness length Z_o. For middle latitudes and more usual extreme gradient wind speeds, equation (13) might be written

$$\frac{V_Z}{V_G} = 0.082\ Z_o^{0.09} \ln \frac{Z}{Z_o} \tag{14}$$

3.2 Advection

In the previous section it was tacitly assumed that the surface of the ground was uniform for a sufficient distance upwind for steady-state conditions to be established. This is obviously not always so and in many instances such as the outskirts of cities the wind profile goes through a transition. An important question is how far down wind of a change in roughness is it before the wind regime appropriate to the new surface is established. Jensen's results give a clear indication of the order of the variation but are insufficiently detailed to be precise.

The problem has however been treated by Taylor[32] who determined approximately the fetch distance downwind. His results are shown in Fig. 10. The parameter r denotes the roughness length ratio $\frac{Z_o^{1}}{Z_o^{11}}$. For a typical example of a wind blowing from flat open terrain ($Z_o^{1} \approx 10$ cm) to a suburban area ($Z_o^{11} \approx 100$ cm), $r = 0.1$. For the wind regime appropriate to a suburban area to establish itself at 100 ft requires a downwind distance of approximately 3/4 mile from the change in roughness. At a height of 300 ft the distance required is about 5 miles. For a wind blowing in the opposite direction (from suburban area to open country) $r = 10$, and the downwind distance required for a 100 ft height is about 2 miles and for 300 ft about 9 miles. Thus it takes almost twice as far for the wind to pick up speed over the smoother surface as it does to slow down over the rougher.

Meteorological stations are in many instances located at airports and the assertion is often made that the wind there is typical of flat open country. The results given above suggest that if the airport is situated on the outskirts of a city for this to be true there should be a clear fetch of some 2 or 3 miles. This is seldom the case even in one direction.

Fig. 10 is clearly of importance in estimating any reduction in mean wind velocity which might be made on account of the rougher nature of the terrain.

3.3 Orographic effects

The orographic effects on the wind (due to mountains and valleys) are extremely difficult to define with generality. The principal effects are the amplification which can arise at mountain tops to the funnelling of wind in valleys. This is generally extremely local and the wind profile and velocity amplification can in most cases only be satisfactorily established by site investigation (see overleaf).

4. CHARACTERISTICS OF GUSTINESS NEAR THE GROUND IN HIGH WINDS

4.1 General

An erroneous convenience commonly adopted in wind loading is that the wind is comparatively steady and hence gives rise to steady pressures. In fact, storm winds are usually extremely unsteady and turbulent: and rates of change of over three hundred miles per hour per second over ranges of 20 or 30 m.p.h. have been measured.

While there is still no completely satisfactory definition of turbulent flow one of its most recognizable features is its randomness. Because of this the description of a turbulent flow reduces to a description of its statistical properties. Typically, a turbulent flow at a point is characterized by a mean velocity and three fluctuating components of velocity in three mutually perpendicular directions. Two types of statistical measure are important. First the probability function defining the distribution of velocities for each component. Second the correlations within the turbulent field between the velocity components. Alternative to and interchangeable with the correlation functions are the spectra and cross-spectra of velocity which define the contributions made to the variances and co-variances of velocity by fluctuations of different frequencies. To completely define the flow through these quantities is a formidable task particularly in the boundary layer where isotropy extends at most to the horizontal. From the practical point of view however, only some of these quantities seem important. These are,

1. The spectrum of horizontal wind speed
2. The spectrum of vertical velocity
3. The spatial correlations of the velocity components at specific frequencies.

These and the probability density are now discussed.

4.2 Probability density

The probability distribution of wind velocity has been examined[3,5,17,28,30] and shown to agree well with the familiar normal or Gaussian distribution given by

$$p(v).\ dv = \frac{1}{2\pi\sigma} \cdot e^{\frac{-(v-\bar{v})^2}{2\sigma^2}} \cdot dv \tag{15}$$

where σ^2 is the variance of the velocity (mean-square fluctuation). From a study of 1 second wind velocity measurements taken over a large number of separate five-minute periods on a 350 ft mast Huss and Portman[18] concluded: "The assumption of a normal frequency curve for representing wind-velocity distribution, while not accurate for any given case, would appear to be justifiable for a large number of cases."

4.3 Spectrum of horizontal gustiness in high winds

The spectrum of horizontal wind speed over an extended frequency range has already been referred to in Fig. 2. Attention is now confined to the high-frequency end of this spectrum pertaining to gusts. From a study of some 90 strong-wind spectra obtained at different heights, surfaces and parts of the world, the writer[10] suggested the following expression for the spectrum of horizontal gustiness.

$$\frac{n.\ S(n)}{V_*^2} = \frac{n.\ S(n)}{\kappa\ \bar{V}^2} = 4\ \frac{x^2}{(1 + x^2)^{4/3}} \tag{16}$$

where $S(n)$ is the spectrum of horizontal speed at frequency n and height z and $x = 4000 \frac{n}{\bar{V}_1}$ where $\frac{n}{\bar{V}_1}$ is in waves per foot.

This curve is illustrated in Fig. 11 together with the averaged results for a number of different localities and heights, some of which were given previously. Additional results are given for the Severn River Railway Bridge[11] and for a 150 ft tower near the centre of London, Ontario. The individual spectra for the latter are given in Fig. 12. The city here is situated in flat country and the tower stands 150 ft high amidst houses and larger buildings up to four stories high along wide avenues lined with trees up to fifty feet high (mainly deciduous). The relationships between drag coefficient and wind profile already outlined suggests that the appropriate value of the drag coefficient κ is 0.030-0.035 and the roughness length Z_0 = 1 m. If so, this site is the roughest for which spectral measurements are available.

It appears that there is quite satisfactory consistency between the various curves for all types of surface and heights above ground (some variation can be attributed to variations in instrument response and method of analysis). Three particular features are noteworthy about the spectrum. First, the peak which occurs at a wavelength of approximately 2000 ft, second the proportionality of the spectrum to $\left(\frac{n}{\bar{V}}\right)^{-2/3}$ for large values of $\frac{n}{\bar{V}}$ (a theoretical result) and third, the proportionality to the shear stress and friction velocity V_*^2. Since the mean-square fluctuation velocity is proportional to the area under the spectrum it follows from that the R.M.S. intensity of turbulence at height Z is

$$I_Z = \frac{\sigma}{\bar{V}_Z} = \frac{\left[\int_o^\infty S(n).\ dn\right]^{1/2}}{\bar{V}_Z}$$

$$= \frac{\sqrt{6 V_*^2}}{\bar{V}_Z} \tag{17}$$

$$= 2.35\sqrt{\kappa}\,\frac{\bar{V}_1}{\bar{V}_Z}$$

4.4 Spectrum of vertical gustiness

The spectrum of vertical gustiness has been studied extensively by Panofsky and McCormick[25]. A suggested empirical formula is

$$\frac{n.\ S_W(n)}{\bar{V}_*^2} = 6\,\frac{f}{(1 + 4f)^2} \tag{18}$$

where $f = \frac{nz}{\bar{V}}$ - the ratio of the height to wavelength. This curve is illustrated in Fig. 13.

Again, a peak is apparent when $\frac{nz}{\bar{V}} = 0.25$, that is, when the wavelength $\frac{\bar{V}}{n}$ is four times the height. An important distinction between vertical and horizontal gustiness is that the former appears to be strongly dependent on the height z.

4.5 Vertical and lateral correlations

When considering the wind loading on an extended structure such as a long bridge, tall mast or skyscraper it is clearly important to have some measure of the spatial distribution of gusts. This can be measured by the correlation coefficient between two velocity measurements spatially separated by the interval Δx. The co-variance between the two velocity measurements is the average product $\overline{V_1\ V_2}$ and the cross-correlation is $\frac{\overline{V_1\ V_2}}{\sqrt{\overline{V_1^2}\,.\,\overline{V_2^2}}}$. Just as the variance could be broken down frequency by frequency into a spectrum, so can the co-variance (or cross-correlation). Following this approach we can define a "narrow band" correlation function or (cross-correlation spectrum) as

$$R_{\Delta x}(n) = \frac{S_{12}(n)}{\sqrt{S_{11}(n) . S_{22}(n)}} \quad (19)$$

Unlike the spectrum itself, the cross spectrum $S_{12}(n)$ can be a complex quantity having both in-phase and quadrature components. The existence of the quadrature component can be taken to indicate a preferred orientation of eddies and therefore only occurs when there is asymmetry present in the flow. For example there is no significant quadrature component in the cross-wind horizontal cross-spectrum between like components of velocity: in the vertical direction however, where there is strong asymmetry, the quadrature component is significant and the maximum correlation between the horizontal wind speed at two different heights occurs not simultaneously, but when the signal from the lower station is delayed by a time roughly equal to $\frac{\Delta z}{\bar{V}}$ where $\bar{V}$ is mean velocity: or where the instruments are positioned on a line inclined at 45° to the downwind direction.

The square of the absolute value of the cross-correlation spectrum is termed the "coherence." For practical purposes it is probably quite adequate to neglect the quadrature correlation, as such, and take the cross correlation as equal to the square root of the coherence.

The general form of the cross-correlation spectrum can best be seen from measurements made in a wind tunnel of the crosswind correlation of wind speed some distance downstream of a plane jet[12]. These are shown in Fig. 14. The cross correlation is expressed as a function of the non-dimensional ratio $\frac{n.\Delta x}{\bar{V}}$. Within the limits of the experimental accuracy the cross correlation can be quite well expressed by

$$R\left(\frac{n.\Delta x}{\bar{V}}\right) = e^{-C\frac{n.\Delta x}{\bar{V}}} \quad (20)$$

where $C \approx 8$

This type of curve appears to be characteristic also of correlations in atmospheric turbulence. Fig. 15 shows measurements of the vertical cross-correlation ($\sqrt{\text{coherence}}$) of horizontal wind speed on masts situated in (a) open grassland, and (b) wooded terrain. Exponential decay curves similar to the above are shown in which the appropriate values of the coefficient C are roughly 7.7 and 6.0 respectively - remarkably close to the value found for the wind tunnel.

The quantity $\frac{1}{C} \cdot \frac{\bar{V}}{n}$ in fact defines the "scale" of the correlation (the distance to the centre of gravity of the correlation diagram or the "effective gust width"). Some valuable measurements of the along-wind and cross-wind scales of the along-wind u and cross-wind velocity components for specific wavelengths have been published by Cramer[7], see Fig. 16. These suggest that the along-wind scales of both the along-wind and cross-wind velocity components are roughly 1/6 of the wavelength in both stable and unstable atmospheric conditions. In unstable conditions the transverse scales of the same two velocity components are again roughly the same but slightly smaller being 1/10 of the wavelength. In stable conditions the cross-wind scales are very much less than the along-wind being roughly 1/40 and 1/25 of the wavelength for the along-wind and cross-wind components respectively.

The indication this gives is that in unstable conditions the along-wind and cross-wind scales are about equal (and equal to about 1/6-1/8 of the wavelength) in stable conditions the eddies are very much elongated in the direction of the wind and the cross-wind scales are of the order of 1/3-1/5 of the along-wind scale which itself is equal to roughly 1/8 of the wavelength. Which of these two models is representative of high winds is not yet clearly established although the weight of evidence is that the elongated eddy is more representative. The strongest evidence for this is probably in the directional traces for strong wind which do not tend to exhibit the wildly meandering characteristics of highly unstable conditions. In the boundary layer of the wind tunnel, Grant[15] who measured all nine correlation functions found that the longitudinal scale was 7 or 8 times larger than the lateral. This of course is also mechanical turbulence, as in strong wind. Some measurements of wind velocity across a broad front were obtained in 1937 by Bailey and Vincent[1] at the site of the Severn River railway bridge. These data were analysed by the writer and have been discussed by Panofsky[24]. From the cross-correlation coefficients between the wind at neighbouring instruments across the span the indications were that the cross-wind scale was somewhat less than 1/3 of the along-wind scale.

In previous papers the writer has suggested that because of the uncertainty, the assumption leading to the higher wind loading should be made and the cross-wind scale taken equal to the along-wind. This may now seem unnecessarily conservative and a somewhat smaller cross-wind scale might be adopted.

To summarize it is suggested that the following may be taken as representative values of the scales of turbulence at specific wavelengths:

Vertical scale of wind-speed	$\frac{1}{7} \cdot \frac{\bar{V}}{n}$
Lateral " "	$\left.\right\}\ \frac{1}{25} \cdot \frac{\bar{V}}{n.}$ (21)
Lateral " " vertical velocity	

4.6 The structure of turbulence

Perhaps the simplest notion concerning the structure of turbulence near a plane but rough surface, is that of a series of "roller type" eddies. This form has been tentatively suggested by Webb[37] for the natural wind and by Townsend[35] for the outer layer in the wind tunnel. Two models along these lines can be envisaged - one in which the eddies at the surface cartwheel along the ground in the direction of the flow and rotate in the same sense and one in which consecutive eddies rotate in opposite senses as indicated by Webb. The former poses the difficulty however that the neighbouring boundaries of the eddies must be moving in opposite directions (one up and one down) with consequent high shear. While the latter could well be typical of convective conditions in which thermal updraft provides the energy for rotating neighbouring cells of air in opposite senses it is difficult to see how such an eddy system could be generated mechanically (as in high wind), since it implies surface motion first opposing the flow then with it. Further it is difficult to see why, even if the turbulence is generated convectively the eddies should prefer to be lined up in the wind direction any more than across wind. In fact the similarity of longitudinal and across-wind scales in unstable conditions (see Fig. 15) suggests they do not have any preferential orientation.

More recently Townsend[34] has quoted some wind-tunnel measurements which are not generally consistent with the "roller" hypotheses. Instead he suggests that the motion consists of two dimensional jets which originate in the immediate neighbourhood of the wall surface. Such jets with their surrounding induced flow are shown diagrammatically in Fig. 17. Within the jet, the streamlines are deflected down wind, by the action of the velocity profile. The slower longitudinal velocity in the jet compared to that outside is due to the transport upwards of the slower moving fluid near the surface. In three dimensions the flow within the jet and the return flow outside resembles in some senses two contra-rotating corkscrews with their axes lying parallel to the mean flow and their inside boundaries both moving upwards, constituting the jet flow.

There are several features of this model which are supported by observations in strong wind. First it accounts for the displacement of the maximum correlation of the longitudinal velocity component in the vertical direction to a position along a line inclined at roughly 45° to the mean flow and coinciding approximately with the mean of the postulated jets. (See discussion on scales.) It accounts for the much larger longitudinal scale compared to the lateral scales. Third it accounts for the influence of height on the vertical scale but its apparently much weaker influence on the horizontal scales. (This is implied by the dependence of the vertical spectrum on the parameter $z\frac{n}{V}$, z being the height, whereas the wind-speed spectrum depends on $L\frac{n}{V}$ where L is a horizontal scale found to be of the order of 4000 ft). Finally this model does not incur high shearing either at the eddy boundaries or at the ground surface - as does the roller model. This in a sense implies that the air flow is taking a path of least resistance or obeying the minimum energy principle.

Before firm theories can be put forward however it would seem desirable to have more information.

5. SYNTHESIS

The discussion of the wind structure so far can be summarized as follows. The wind is caused by the presence of large weather-map-scale disturbances which give rise to air movements having characteristic periods of the order of hours and velocities given by the gradient (or geostrophic) velocity. Nearer the ground the roughness of the surface sets up shear forces between the air and the ground which retard the flow throughout the planetary boundary layer. The mean-wind profile in this layer, which approaches the gradient-wind velocity at the top of the layer, can be expressed by power law or other profiles which only depend significantly on the characteristic roughness of the surface. The surface wind speed depends on the roughness and can vary greatly over short distances. In strong winds the boundary layer is fully turbulent. The shear stress at the surface induces mechanical turbulence the energy of which is proportional to the shear stress and depends significantly on the roughness of the surface and the height above the surface. The characteristic frequencies of the mechanical turbulence or gusts are altogether different to those of the weather-map fluctuations and a gap appears to exist in the spectrum of wind speed centered at a period of roughly one hour. This justifies the use of mean hourly wind speed, or somewhat shorter averaging periods as the characteristic velocity. The wind speed and vertical gust spectra can be expressed to an adequate accuracy knowing only the mean hourly velocity and the roughness of the surface.

All of this boils down to the simple fact that almost all the properties of the wind that might be needed in structural design can be estimated reasonably accurately provided that the mean-wind field and the ground roughness are known.

5.1 The mean-wind field

The properties of the mean-wind field of major interest in structural engineering relate mainly to the extremes and their recurrence intervals. Another property, the overall statistical distribution defining the proportion of the total time the wind lies between certain values would also be needed in problems dealing with fatigue (also wind power), but it may be several years before such information is generally applied. (Some illustrations of these distributions have been given by Tagg[31].

There would seem to be two approaches in defining the wind field, one is to define the surface wind field itself, the other, to define the gradient wind and from it estimate the surface wind from the roughness and the wind profile. Whichever method is used - and the relative merits will be discussed below - it seems that the only suitable data on which the wind field can be based are surface measurements from meteorological stations. Another possibility - the use of isobaric maps - has been investigated by the writer, but does not appear promising. These are only available two or three times a day, and so do not constitute a continuous record. They are inevitably to some extent subjective - depending on the meteorologist who draws the map - and furthermore they are least accurate when the winds are strongest and the isobars.most closely packed.

If the wind field is to be based on surface measurements then it may seem most logical to define it also in terms of a surface velocity at some arbitrary reference height rather than the gradient velocity. A difficulty that this approach presents is that, in general, changes in surface roughness take place on a far smaller scale than the grid spacing of meteorological stations (as Jensen's results well indicate). Interpolations are still needed therefore if adequate estimates are to be provided at all localities. A further difficulty lies in the reliability of the individual surface estimates themselves. The periods for which meteorological records are available varies widely; the heights at which the instruments are established vary; frequently they have been moved up and down and from place to place to make room for new buildings and in so doing the exposure has been changed; urban sprawl may have appreciably changed the effective roughness of the surroundings; at coasts the instruments are situated in regions of wind transition and the records are not typical of wind conditions a short distance inland, unserviceability may have broken the continuity of the records. For all these reasons some basis of comparability and cross checking of records seems essential. The most obvious basis of such a comparison is the estimate of the gradient wind made from surface records.

Unlike surface winds, the scale of variations of gradient wind are very much larger than the normal grid of meteorological stations and estimates from several stations should normally overlap. This, in fact, is an argument in favour of using the gradient wind to specify the mean-wind field - at least as an intermediate step. Whether or not it is finally stated in this form for use by structural engineers is another matter. Possibly the gradient wind is too nebulous a concept for down-to-earth engineers! On the other hand the fact that specification in terms of the gradient wind would require the engineer's participation in evaluating the exposure of the structure, and hence the wind loading itself may be of direct benefit in eventually arriving at realistic wind loading.

5.2 Extreme winds

The recurrence of extreme wind speeds can be estimated using extreme value statistics whereby the distribution of largest values in samples of given size can be estimated. It has been indicated above that the wind speed averaged over 1 hour (or thereabouts) is a key quantity in estimating wind loads. If so, what are needed are estimates of the probability of given values of hourly velocity occurring during periods commensurate with the lifetime of the structure. Another way of expressing this is to determine the recurrence interval for different values of hourly wind velocity. A useful time interval is the year. The application of extreme value analysis to wind loading has been discussed elsewhere by Court[8], Shellard[27], Gumbel[16], Thom[33], Jenkinson[19], Boyd and Kendall[4] and the writer[9]. Boyd and Kendall compared three different extreme value methods as applied to wind speed (the Fisher-Tippet 1 curve as used by Gumbel, the Fisher-Tippet 2 curve as used by Thom and A. F. Jenkinson's combined approach). The three methods differ mainly with regard the assumptions on the lower bound and sample size. They are compared with the observed distribution of extreme winds over 32 year period at Victoria, B.C. in Fig. 19. Boyd and Kendall conclude that although none of the methods is ideal for small samples, there is little reason for preferring the methods used by Thom or Jenkinson to the more straightforward method developed by Gumbel. Court analyzed the records of twenty-five weather stations in the United States having 37 years of satisfactory records by the Gumbel method and stated "all of the wind data seems to follow the theory."

The distribution function is of the form

$$F(x) = e^{-e^{-y}} \qquad (22)$$

where y is the reduced variate given by

$$y = a\,(x - U) \qquad (23)$$

a being the dispersion factor and U the mode. For return periods (r) greater than about 10 years, the corresponding value of x is

$$x \approx U + \frac{1}{a} \ln r \qquad (24)$$

The significant factors in this expression are a and U; knowing these, the wind velocity having any desired return period may be estimated. The territorial variation of U and $\frac{1}{a}$ can in a sense be taken as defining the extreme-mean-wind-velocity field. The U and $\frac{1}{a}$ values defining the extreme-gradient-velocity field can be estimated from surface values using these relationships such as (8) and (14) provided the ratio of surface to gradient wind speed can be assessed. In an earlier paper the writer[9] attempted this for the British Isles using values of U and $\frac{1}{a}$ estimated by Shellard[27] from surface records and estimating the ratio of surface to gradient wind speeds using estimates of ground roughness from descriptions of the stations and their surroundings. The results were reasonably encouraging but it was felt that much better could have been done if site inspection and field tests could have been made to determine the roughness more directly. A modified form of the original map is shown in Fig. 19. The surface wind is then determined from the wind profile most appropriate to the terrain and corresponding to the found value for the gradient wind.

The method used in allowing for the variations in surface roughness can be assessed from a comparison of extreme wind speeds at city and airport offices for various cities in the United States shown in Table 1. The data was kindly made available to the writer by Mr. H. C. S. Thom of the U.S. Weather Bureau. The wind speeds refer to the 2 per cent quantile (once in fifty years). As can be seen, in spite of the much greater anemometer height Z_A in the city the wind speed is substantially less (the average being about 70% of the airport speed). The actual ratios are seen to give not unreasonable agreement with the predictions made according to the earlier paper[9].

5.3 Site investigation

It is normal engineering practice for the design of important structures to be preceded by investigations of the various geophysical properties of the site, in particular the geology, the soil and ground water. The value of having detailed information on these pays for the cost of such investigation many times over. Site investigation of the wind conditions could prove to be no less useful and could be carried out at very little expense. The main purpose of such an investigation would be to determine the wind profile - and hence the ratio of surface to gradient winds - the roughness and the gustiness. They would be particularly valuable at hill-top sites where the more usual aerodynamic profiles may be considerably distorted. Such investigations have in fact been conducted in connection with wind-power sites.

The equipment needed would amount to no more than a portable mast with anemometers and field recorders, and pilot balloon equipment or smoke trail rockets would be adequate for most purposes. Records of strong wind of 15-20 miles/hour on two or three occasions would probably provide most of the information that a more protracted study would yield.

TABLE 1

COMPARISON OF EXTREME WIND SPEEDS
AT CITY AND AIRPORT STATIONS IN THE UNITED STATES

	City Office			Airport			City Office		Airport		Ratio: City/Airport	
Station	Z_A (feet) [1]	2% quant. @ 30 ft m.p.h. [1]	2% quant. @ Z_A m.p.h. [2]	Z_A (feet) [1]	2% quant. @ 30 ft m.p.h. [1]	2% quant. @ Z_A m.p.h. [2]	Roughness C'tgy [3]	K_A (C.O.) [4]	Roughness C'tgy [3]	K_A (A.P.) [4]	Actual	Predicted [5]
Boston	188	55	72	63	93	103	8	2.3	3	1.5	0.70	0.65
New Haven	155	47	60	42	70	74	7-8	1.85	4	1.81	0.81	0.98
Chicago		46	57	38	68	70	(Roughness difficult to estimate)					
S. S. Marie	52	54	63	33	84	85	7	3.0	4-5	2.1	0.74	0.70
Kansas City	181	48	63	76	83	95	7-8	2.3	3	1.5	0.66	0.65
Omaha	121	53	65	68	81	91	7-8	2.6	4	1.70	0.71	0.65
Knoxville	111	47	57	71	79	89	7-8	2.6	5	1.90	0.64	0.73
Nashville	191	55	73	42	82	86	7-8	2.3	4-5	1.95	0.85	0.85
Spokane	110	42	51	29	78	78	7-8	2.6	3	1.75	0.65	0.67

NOTES

1. From data provided by U. S. Weather Bureau.
2. Readjusted using 1/7th power law (see Ref. 1).
3. Estimated roughness category of anemometer environment (2-3 mile radius) in direction of prevailing wind from U.S. Weather Bureau descriptions using Table 2 of Ref. 2.
4. Ratio gradient to surface wind velocity (anemometer height) from Fig. 4, Ref. 2.
5. Ratio K_A (A.P.) to K_A (C.O.).

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6. CONCLUSIONS

This paper has attempted to describe as far as possible the macro- and micrometeorological structure of strong winds in the earth's boundary layer insofar as it affects the wind loads on structures. It seems appropriate to conclude by referring to some of the practical implications.

This study has shown that extreme mean wind speeds at building height (say 30 ft) can vary widely. As well as the large-scale global variations in wind climate, much more local variations are also important. It has been shown that the roughness of an urban area itself may slow the wind down to a third of the speed in open country at the outskirts. The corresponding decrease in mean wind pressure (proportional to the velocity squared) is one ninth. In addition, due to different return periods for strong winds, it may be logical to design a monumental structure for a mean wind velocity up to 1½ times greater than that for a shorter life structure, corresponding to a mean wind pressure up to twice as large. These two influences alone (roughness and lifetime) call for a possible factor of 18 in the variation of mean wind loads. To this must be added (or subtracted) the effect of gusts. The wind in a city is much gustier than in open country (by a factor of three or four). Whether this is significant or not depends on the susceptibility of the structure to gusts, that is, whether the natural frequency of the structure coincides with the frequency of high energy fluctuations in the wind spectrum, whether the damping is high or low, and whether the scale of the gusts at critical frequencies is large or small compared to the structure. These considerations taken together suggest a range of values for a logically consistent and realistic set of wind pressure standards which at first sight may seem staggering and is in contrast to the blanket wind loading currently used.

It is fair to ask whether these variations are economically significant or only marginal. The answer is probably that a careful tailoring of wind loading would in many operations such as power transmission, long span bridging and tall skyscrapers and towers provide very substantial savings. Very often in such structures the benefits from small decreases in wind loads snowball rapidly. Smaller structural members are needed, often implying less exposed area to the wind and again smaller wind loads, implying lighter structures and less expensive foundations etc. etc.

The economic value of site investigations in achieving some of these possible savings still has to be fully appreciated. Moreover the savings (or additional utility) that a more careful statement of wind loads would yield, would pay for the necessary exploration and research countless times over.

What research seems necessary? Fundamental to the entire subject of wind loading is of course the careful mapping of the large-scale territorial wind field. Work on the estimation of extreme wind speed parameters at individual weather stations is a valuable step: much needs to be done however in the evaluation of the quality of the measurements themselves. A method for integrating and correlating the measurements into a wind field is suggested by the map given of extreme-gradient-wind parameters. This could be improved probably in a variety of ways, but principally by simple soundings of the wind profile at weather stations (for example by instrument tower, balloon or smoke rocket).

As suggested, the most useful parameter on which to base the wind field is the mean velocity taken over a period between five minutes and one hour (the difference between the two means is not likely to be great). Other measurements such as maximum gust speed are not highly meaningful quantities in relation to wind loads.

More detailed research still needs to be done in determining the wind structure over an urban area - its profile and spectrum. Some general work on spectra and scales of turbulence is in progress at a number of places and this may give rise to modifications to the values suggested here - but the consistent trend in the work so far suggests these may not be large.

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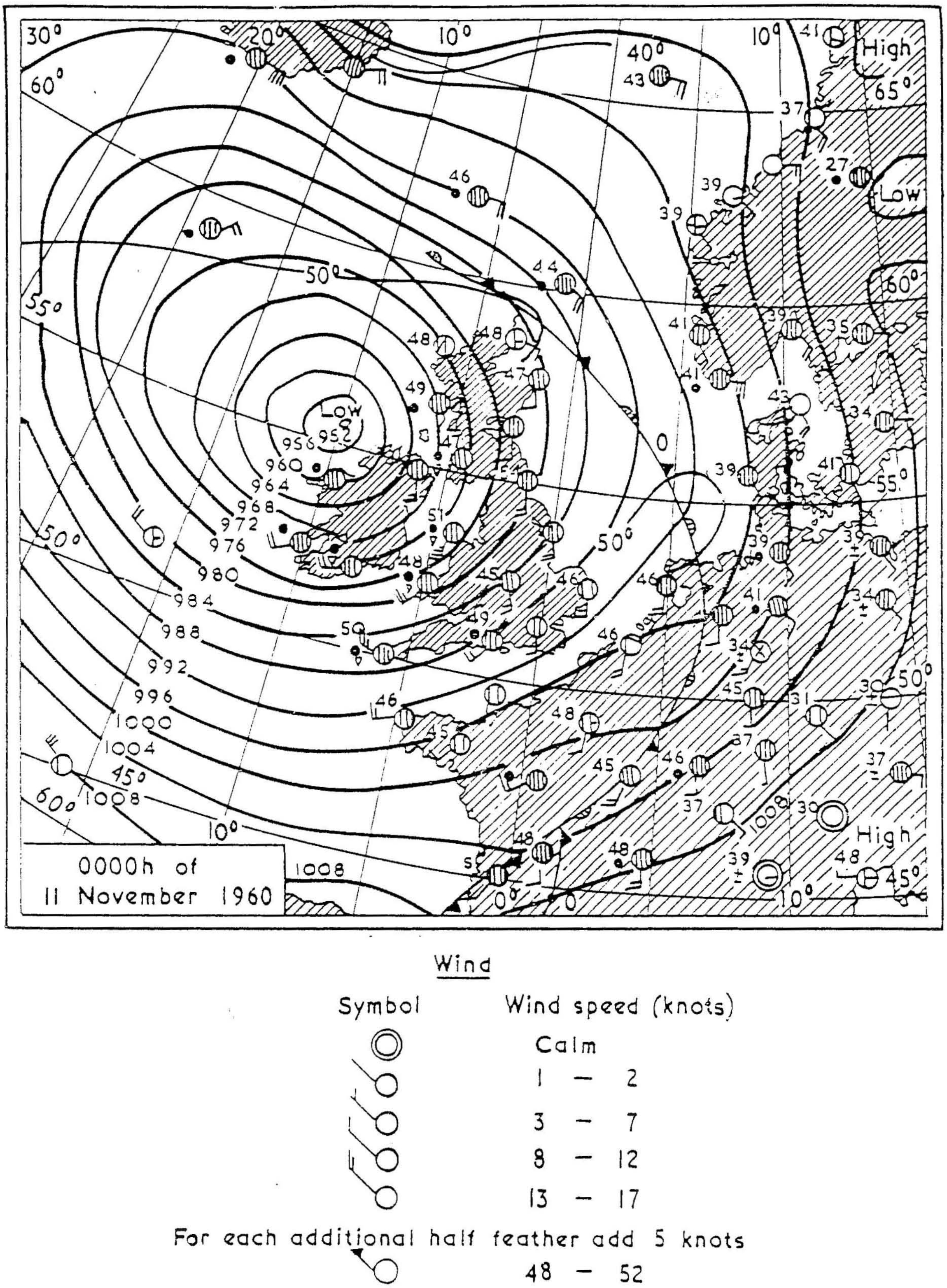


Fig.1. Weather map showing an intense depression with strong winds.

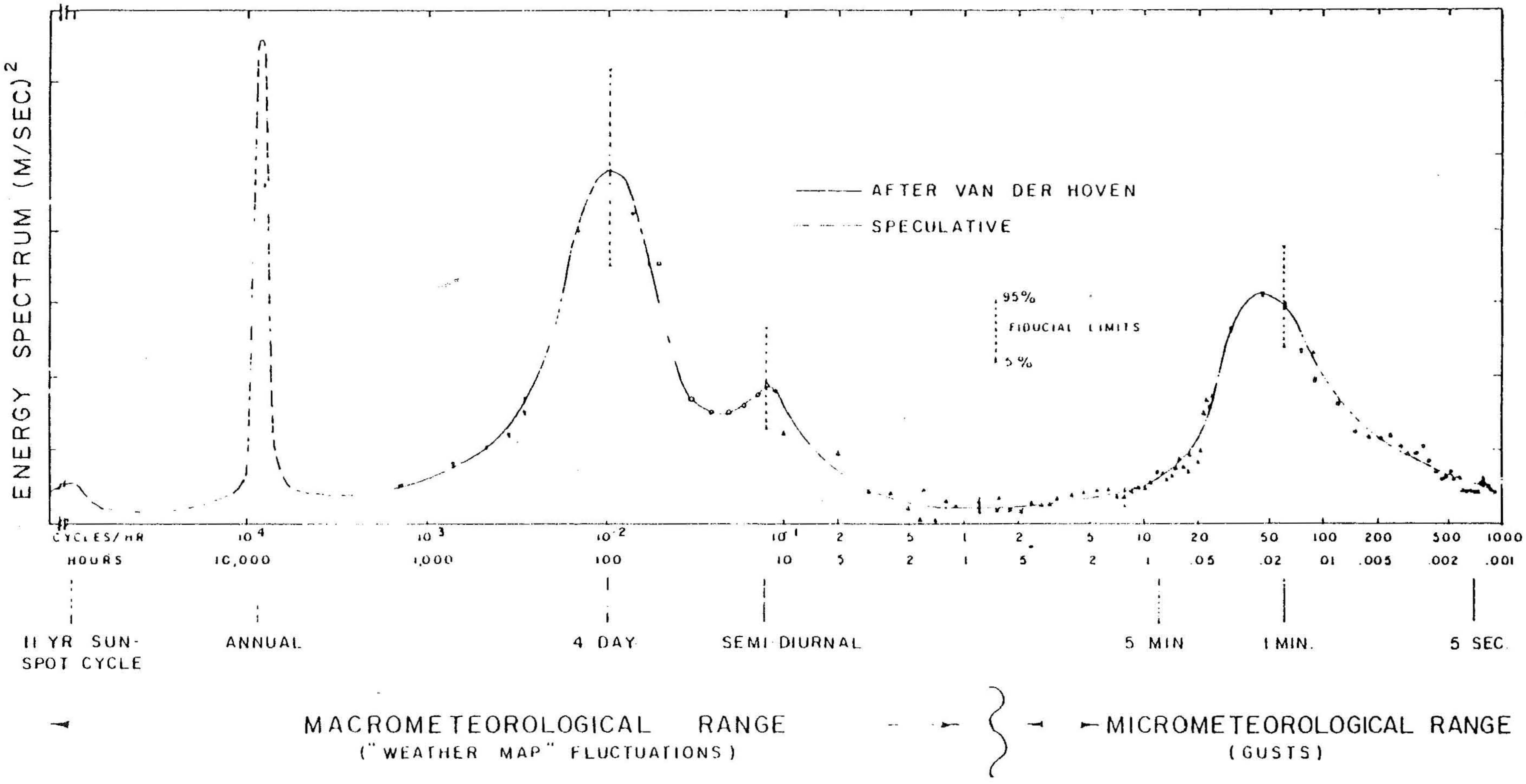


Fig.2. Spectrum of horizontal wind speed near the ground for an extensive frequency range (from measurements at 100 metre height by Van der Hoven at Brookhaven, N.Y., U.S.A.)

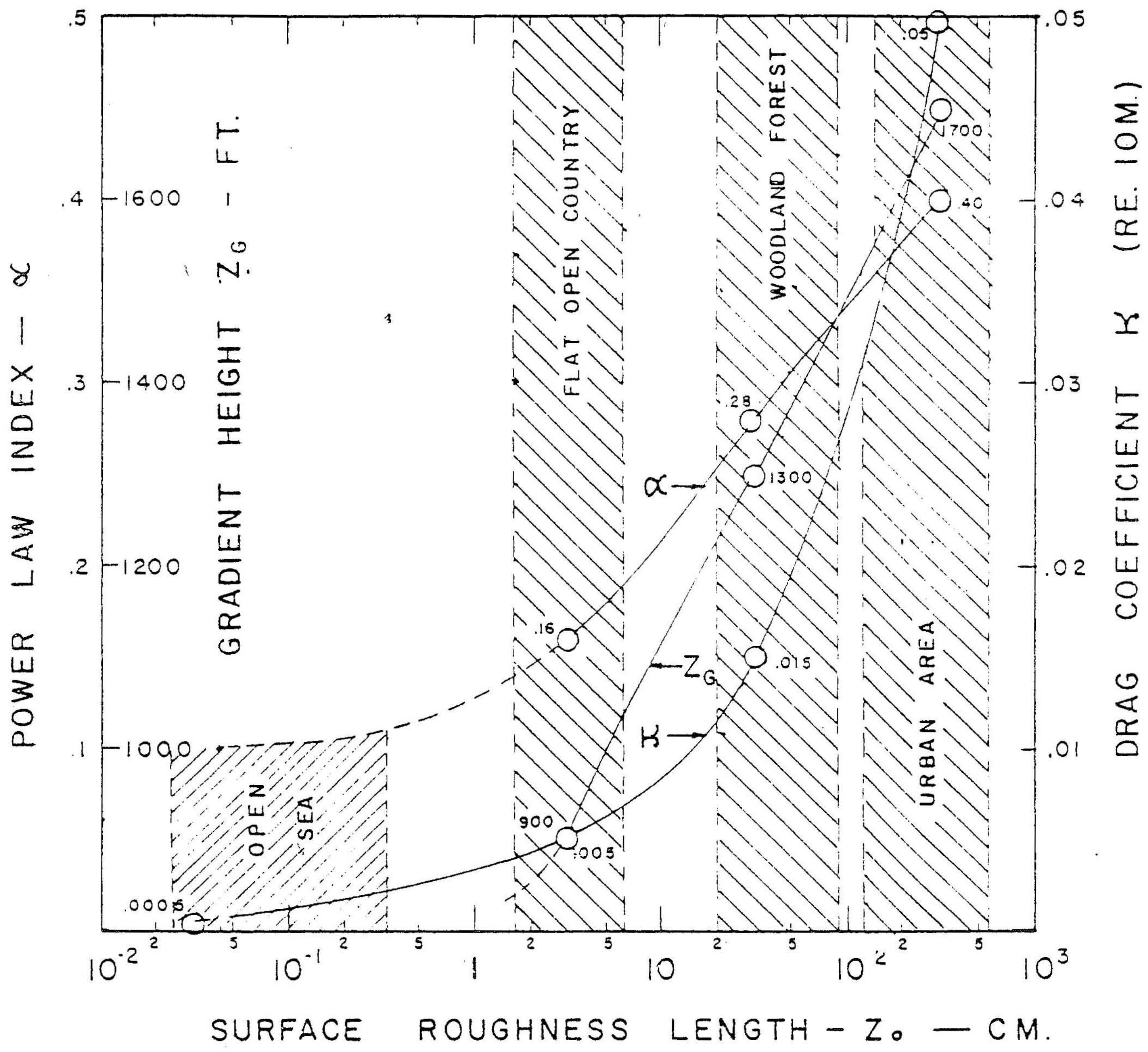


Fig.3. Parameters of wind profile for different surfaces.

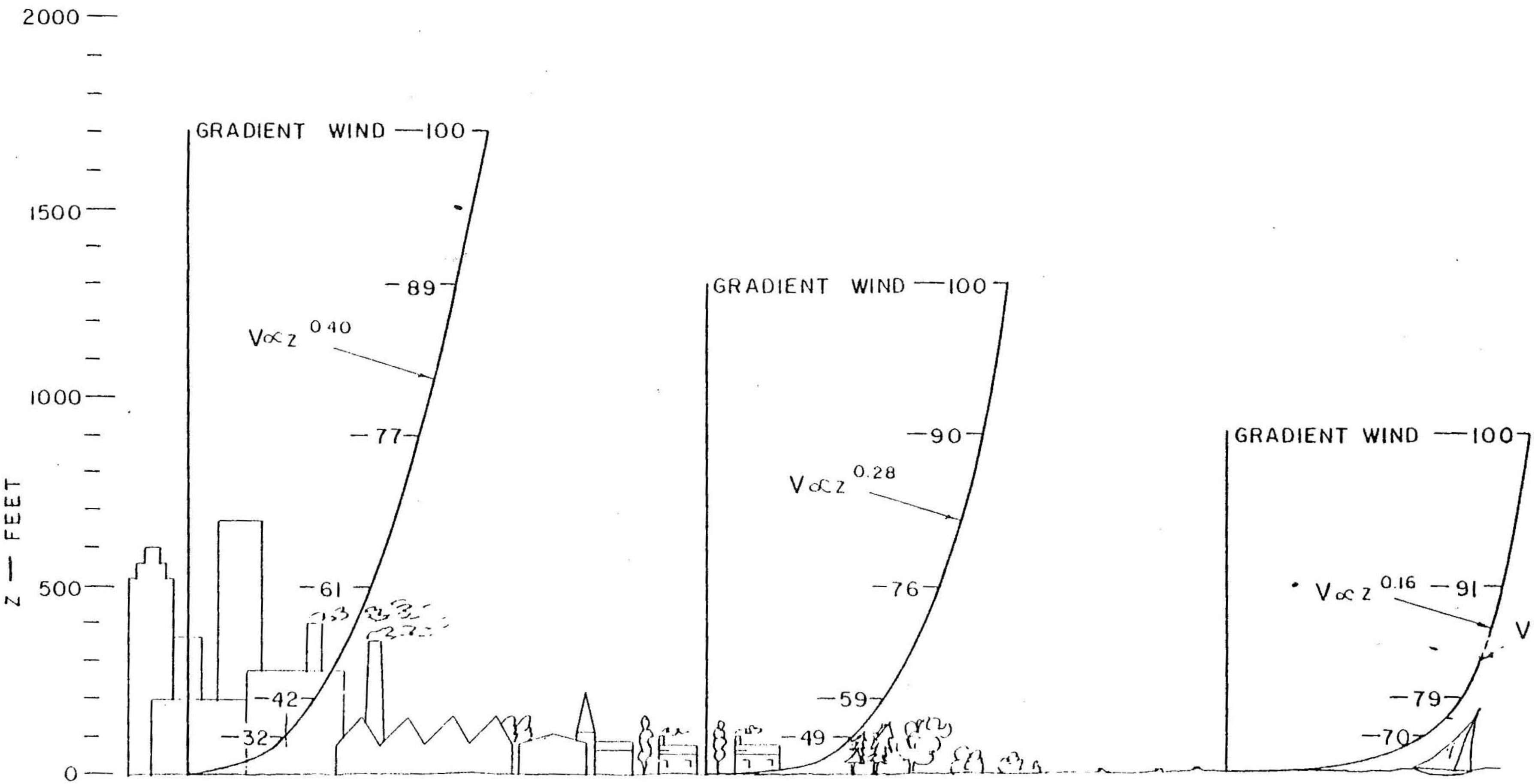


Fig.4. Profiles of mean wind velocity over level terrains of differing roughness.

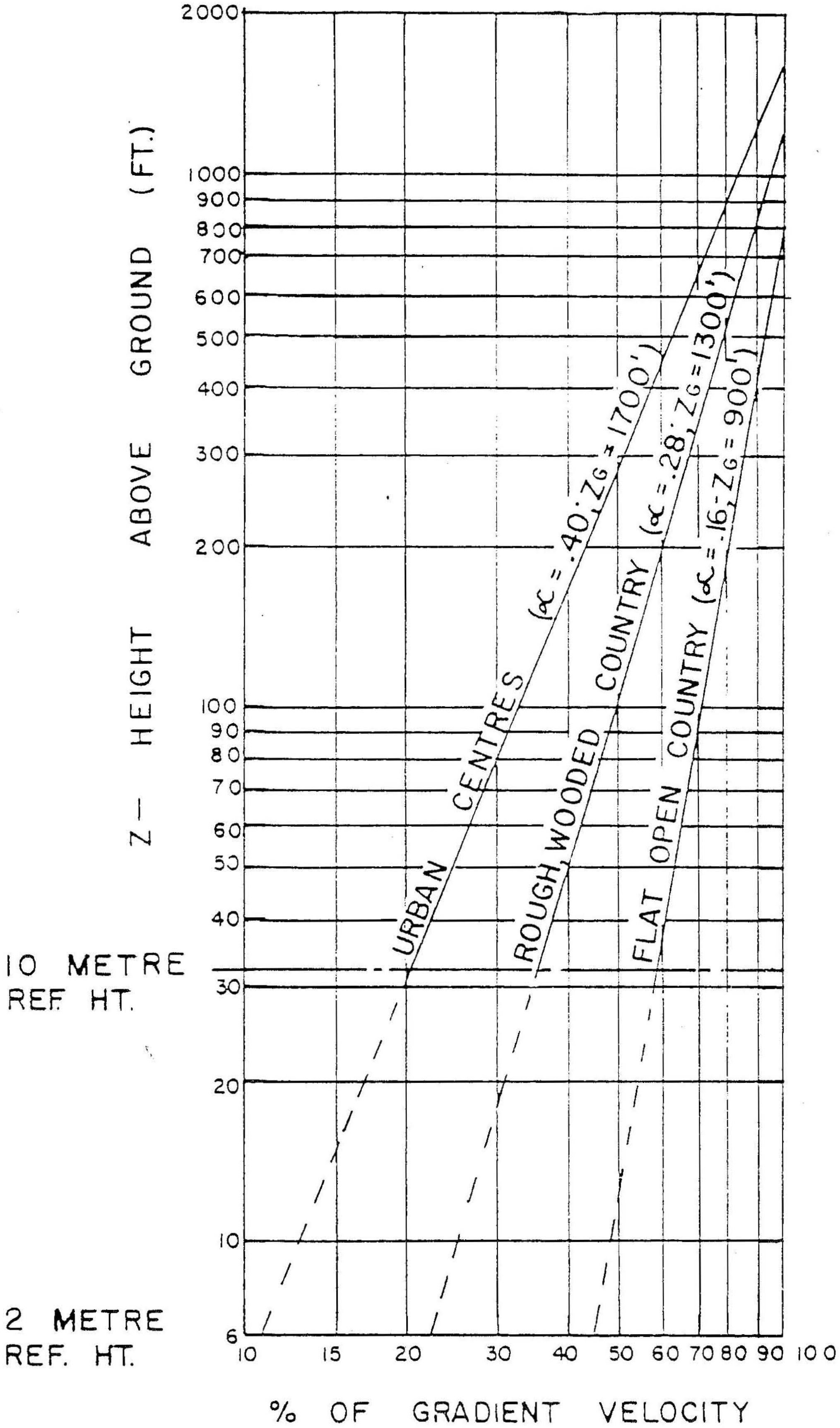


Fig.5. Power law wind velocity profiles for surfaces of different roughness.

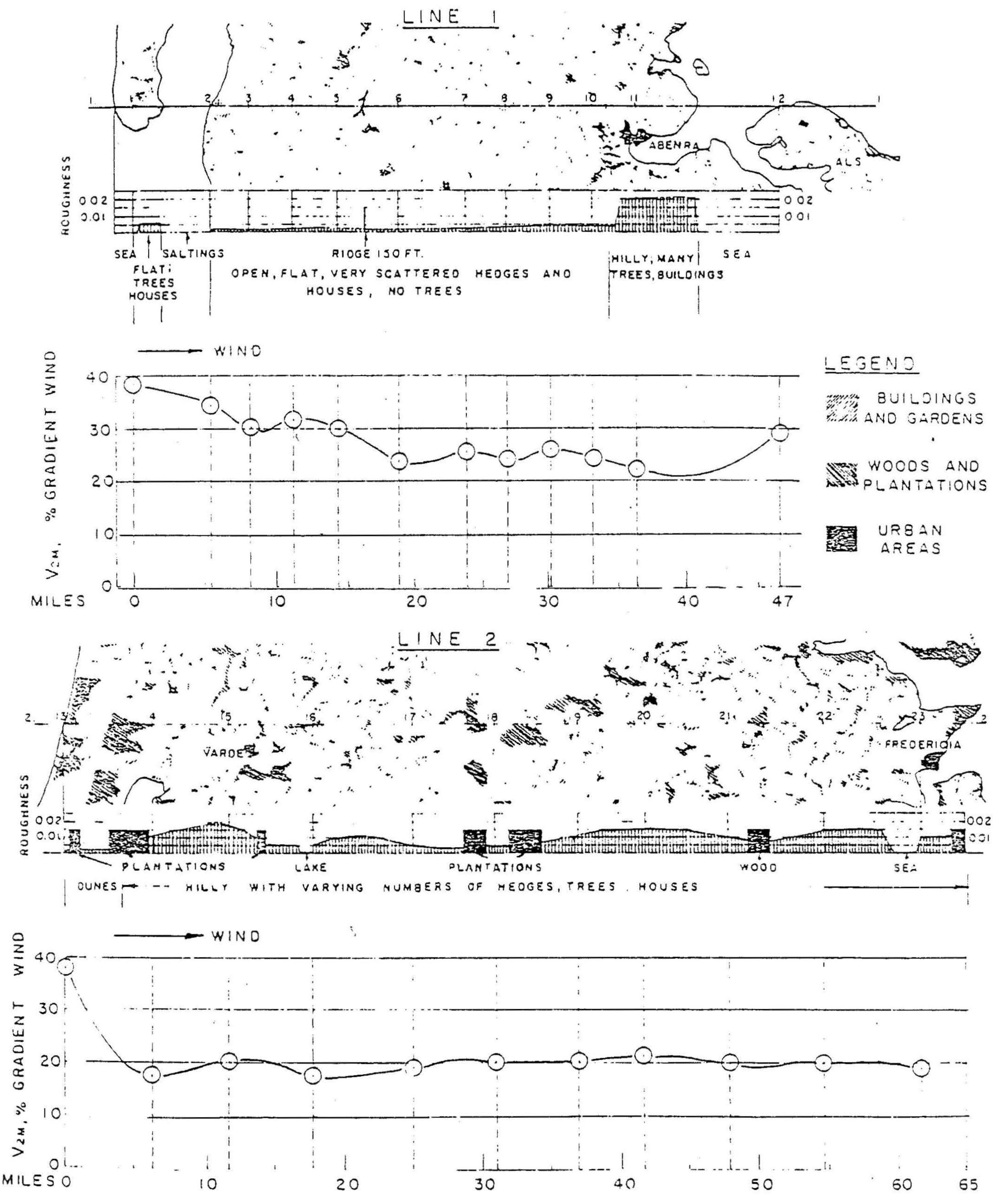


Fig.6. Jensen's measurements of the large scale variations in surface wind velocities with ground roughness (as indicated by the roughness of great order).

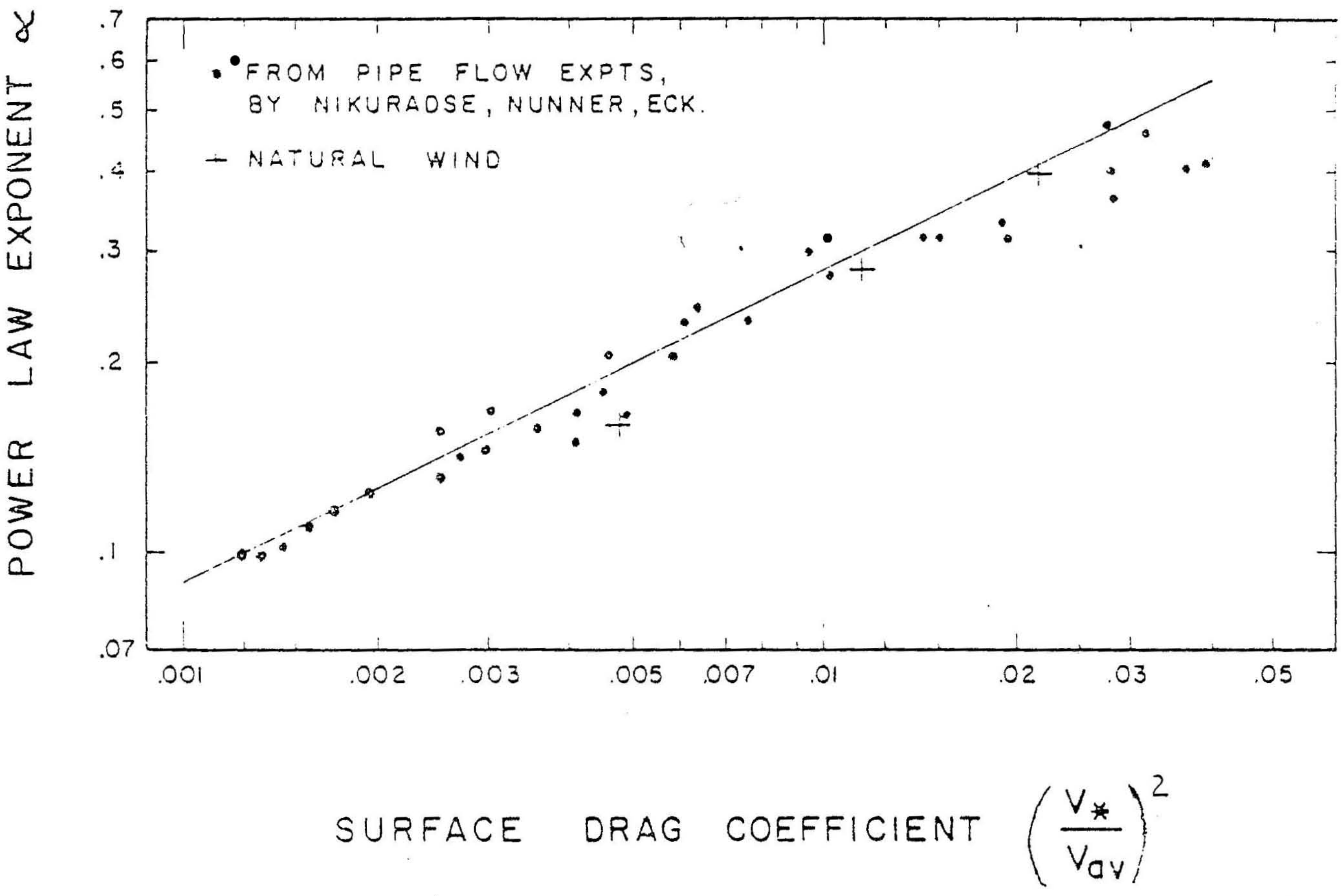
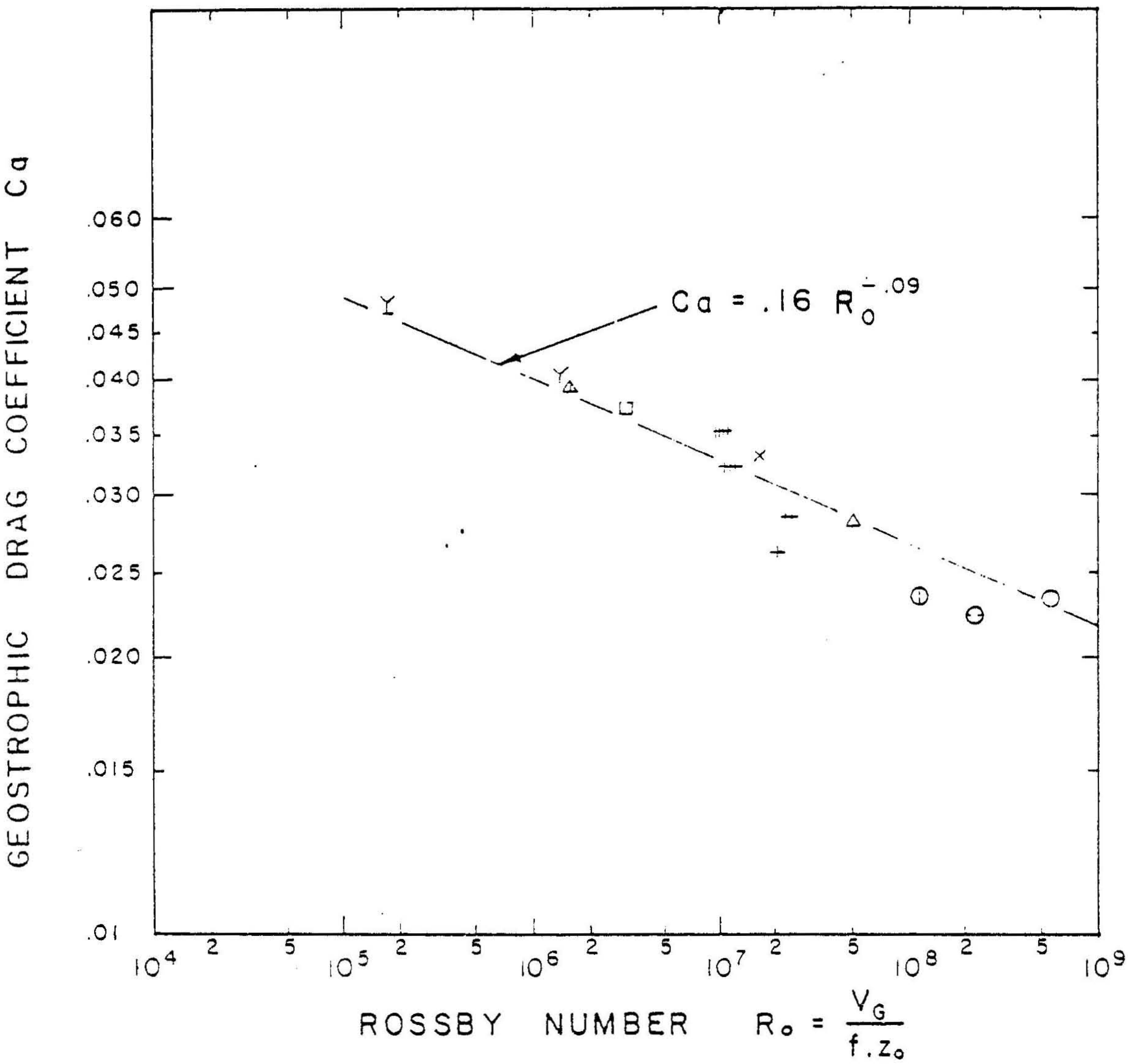


Fig.7. Comparison of turbulent boundary layer parameters in pipeflow and for wind over natural surfaces.



LEGEND (AFTER LETTAU)		LAT.	Z_o (CM.)	$V_{g,o}$ (CM./SEC.)
△	QUICKBORN	53.8	(5)	1000
△	CAMBRIDGE	52.2	0.15	900
□	LEIPZIG	51.4	(5)	1750
⊖	SCILLY IS. I	49.9	(0.05)	1220
○	" " II	49.9	(0.05)	660
Y	MUNICH FOREST I	47.9	20	500
Y	" " II	47.9	10	1500
×	O'NEILL	42.5	0.9	1450
+	COLLEGE STATION I	30.5	0.3	1000
+	" " II	30.5	0.6	1000
+	" " III	30.5	1.8	1500
+	" " IV	30.5	3.3	2400
○	JOHNSON ISLAND	16.5	(0.05)	1160

Fig.8. Lettau's geostrophic drag coefficient as a function of Rossby number.

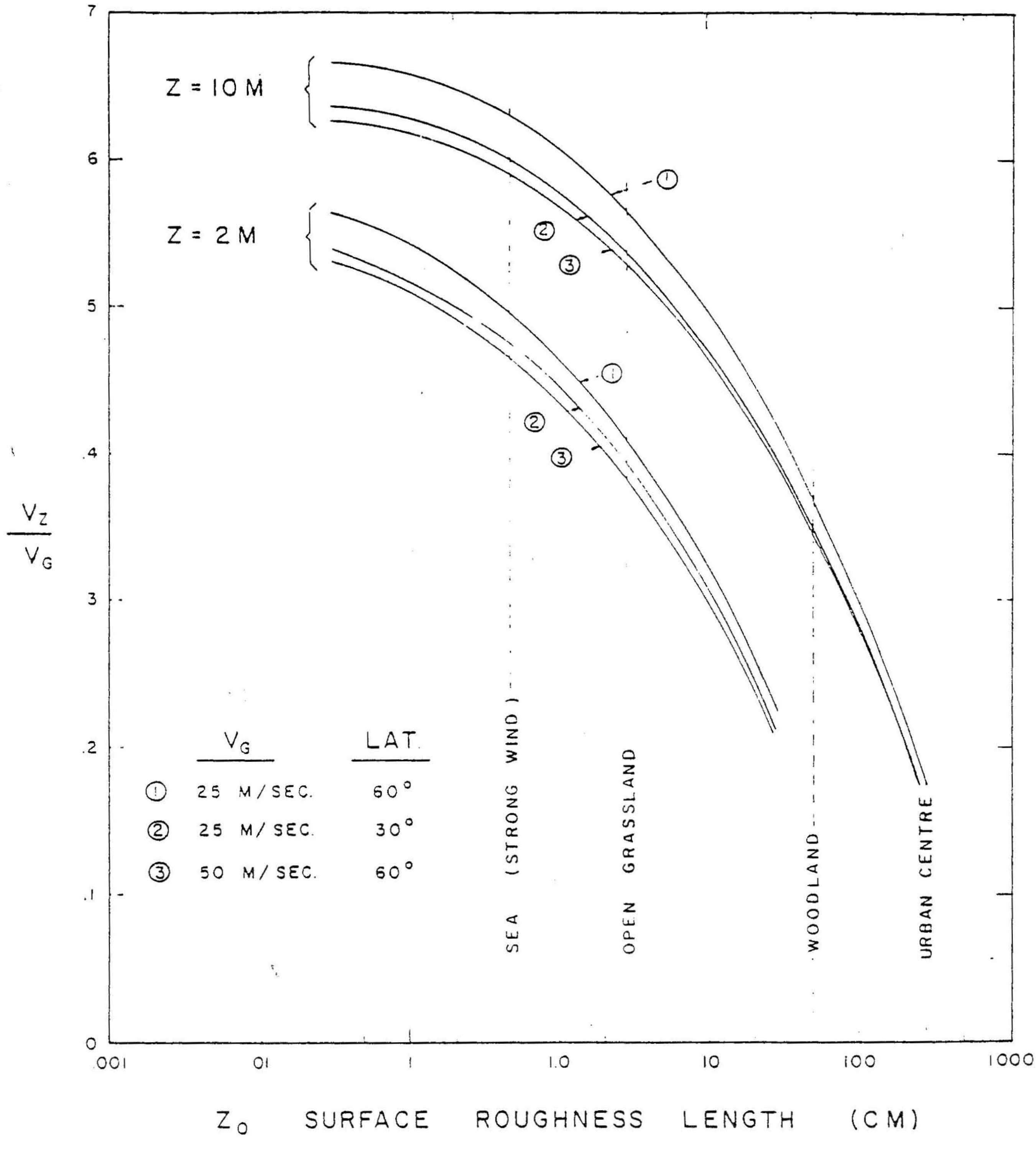
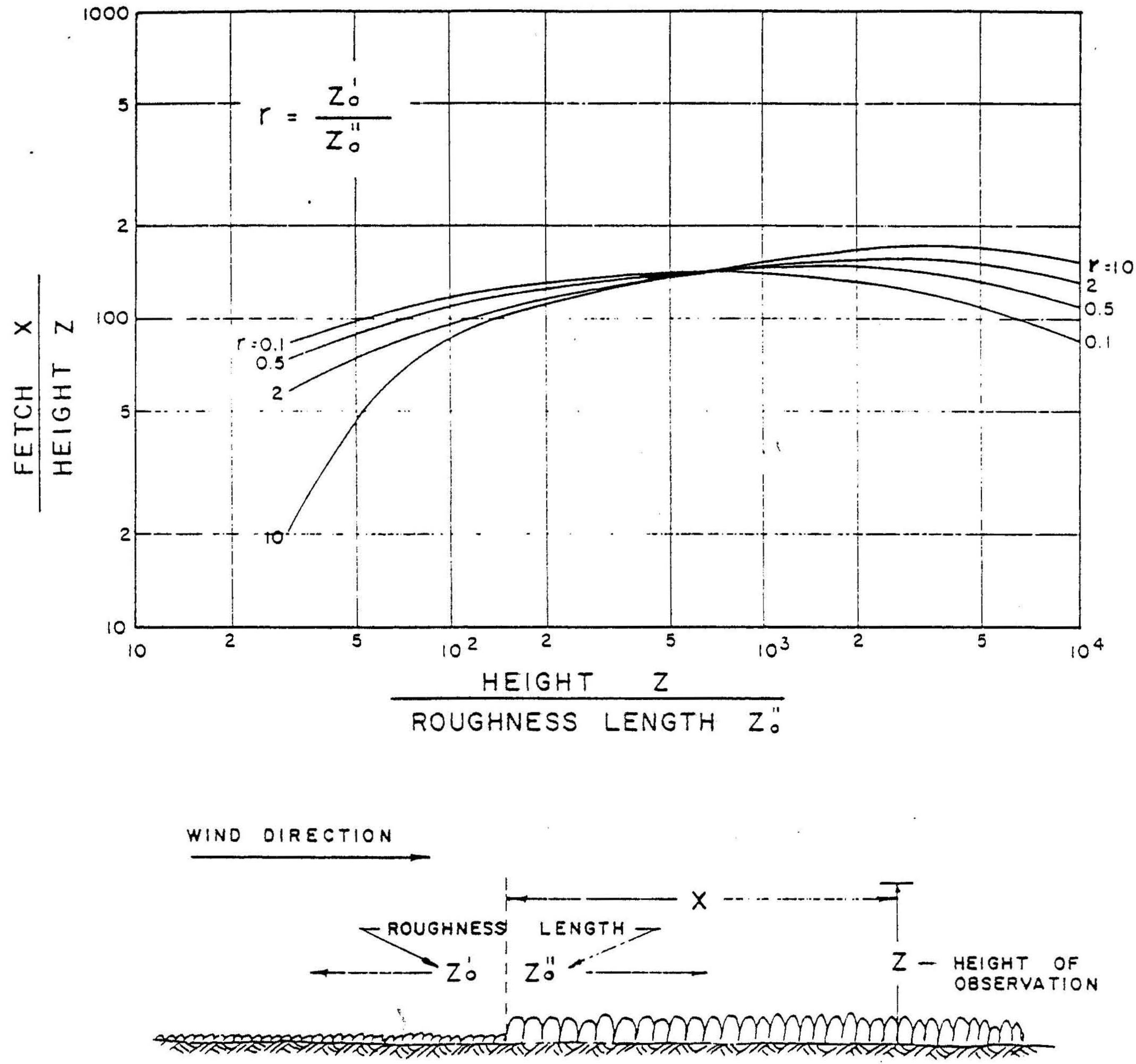


Fig.9. Ratio of surface wind speed to gradient wind speed as function of roughness length.



X = FETCH DISTANCE DOWNWIND OF CHANGE IN ROUGHNESS NECESSARY FOR ESTABLISHMENT OF WIND PROFILE APPROPRIATE TO NEW ROUGHNESS UP TO HEIGHT OF OBSERVATION.

Fig.10. Estimates of fetch distance necessary to establish new wind profile after change in roughness (after Taylor).

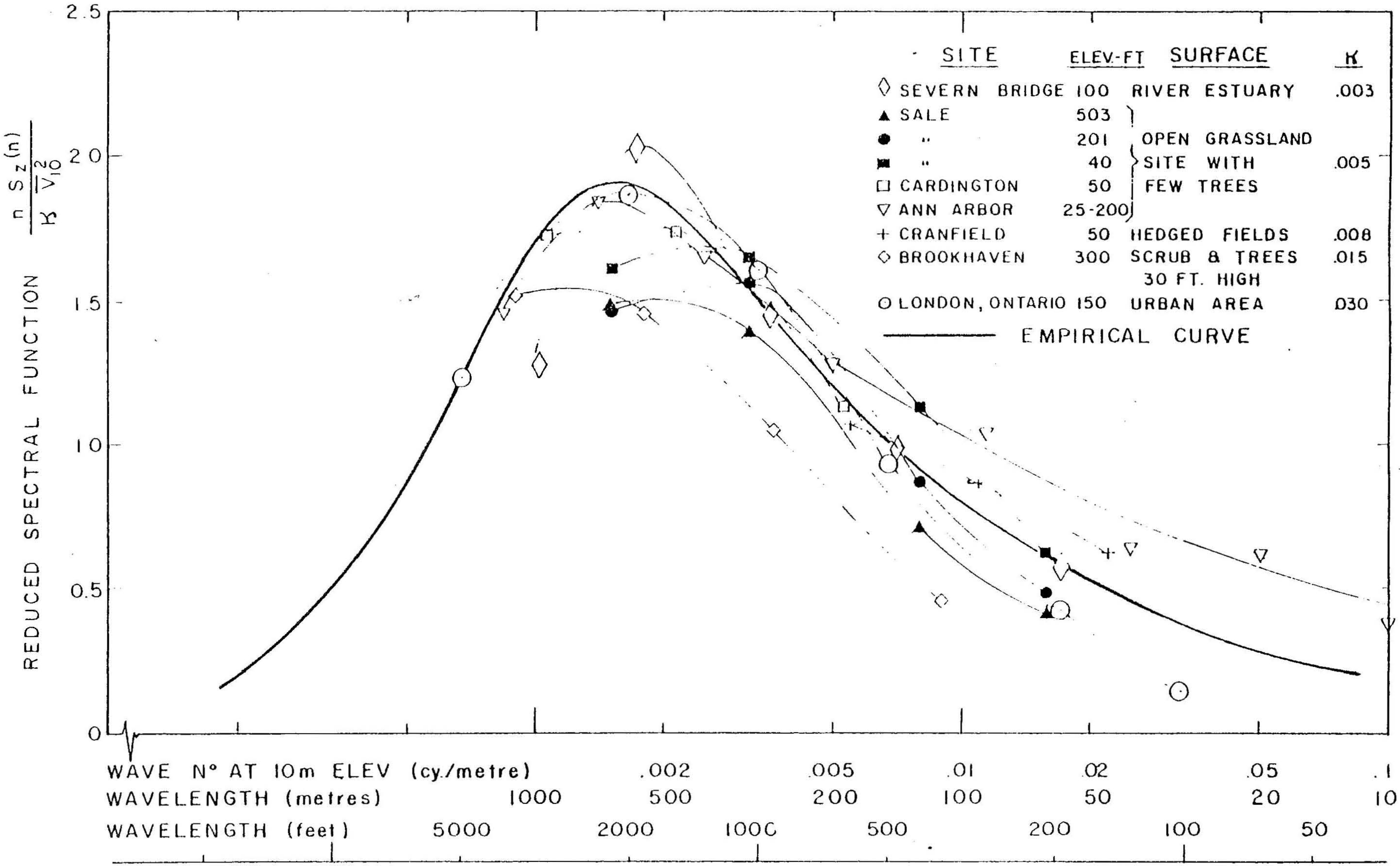


Fig. 11. Spectrum of horizontal gustiness in high winds.

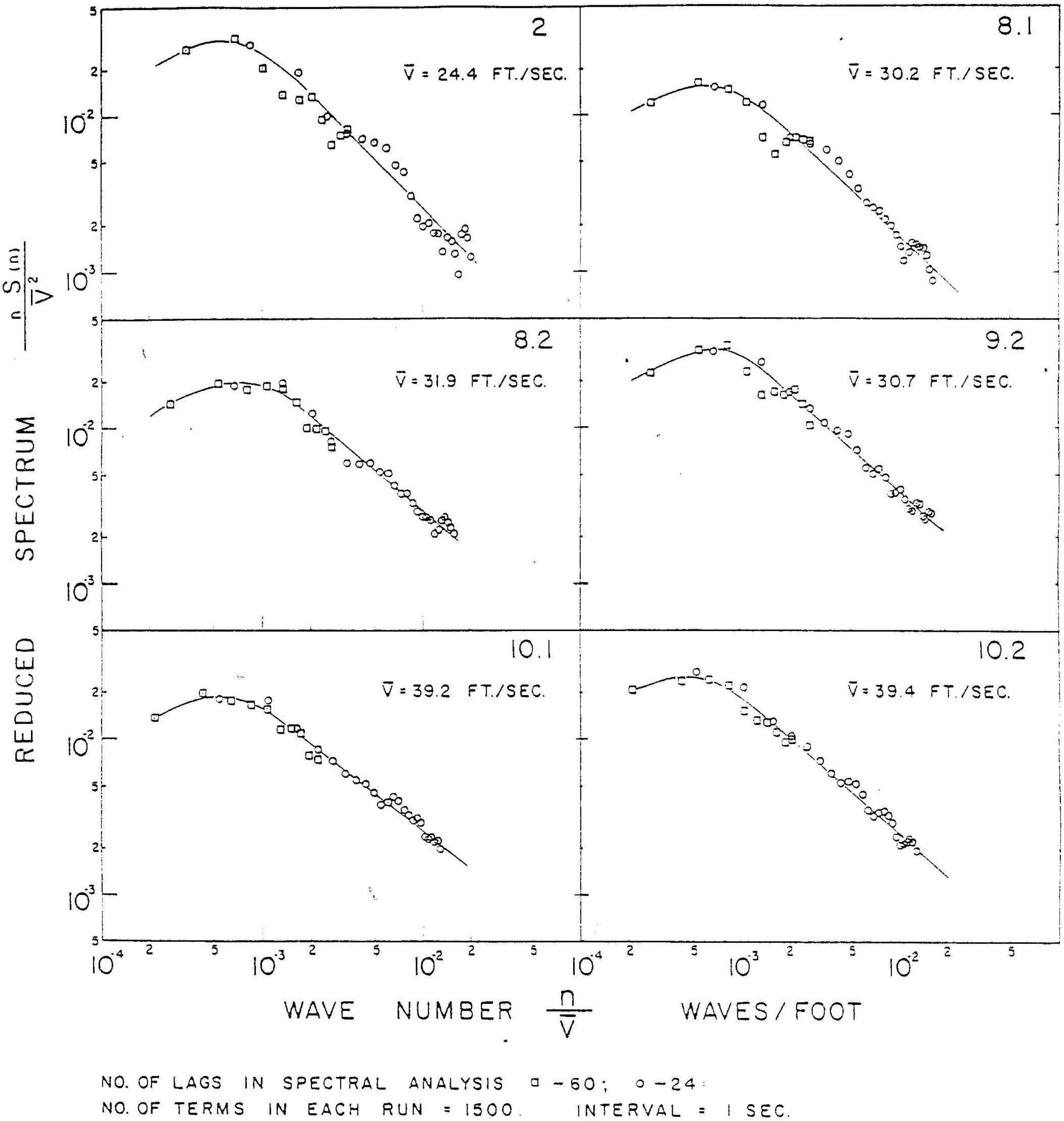


Fig.12. Spectra of horizontal wind speed at 150 ft. in strong wind over urban area (London, Ontario).

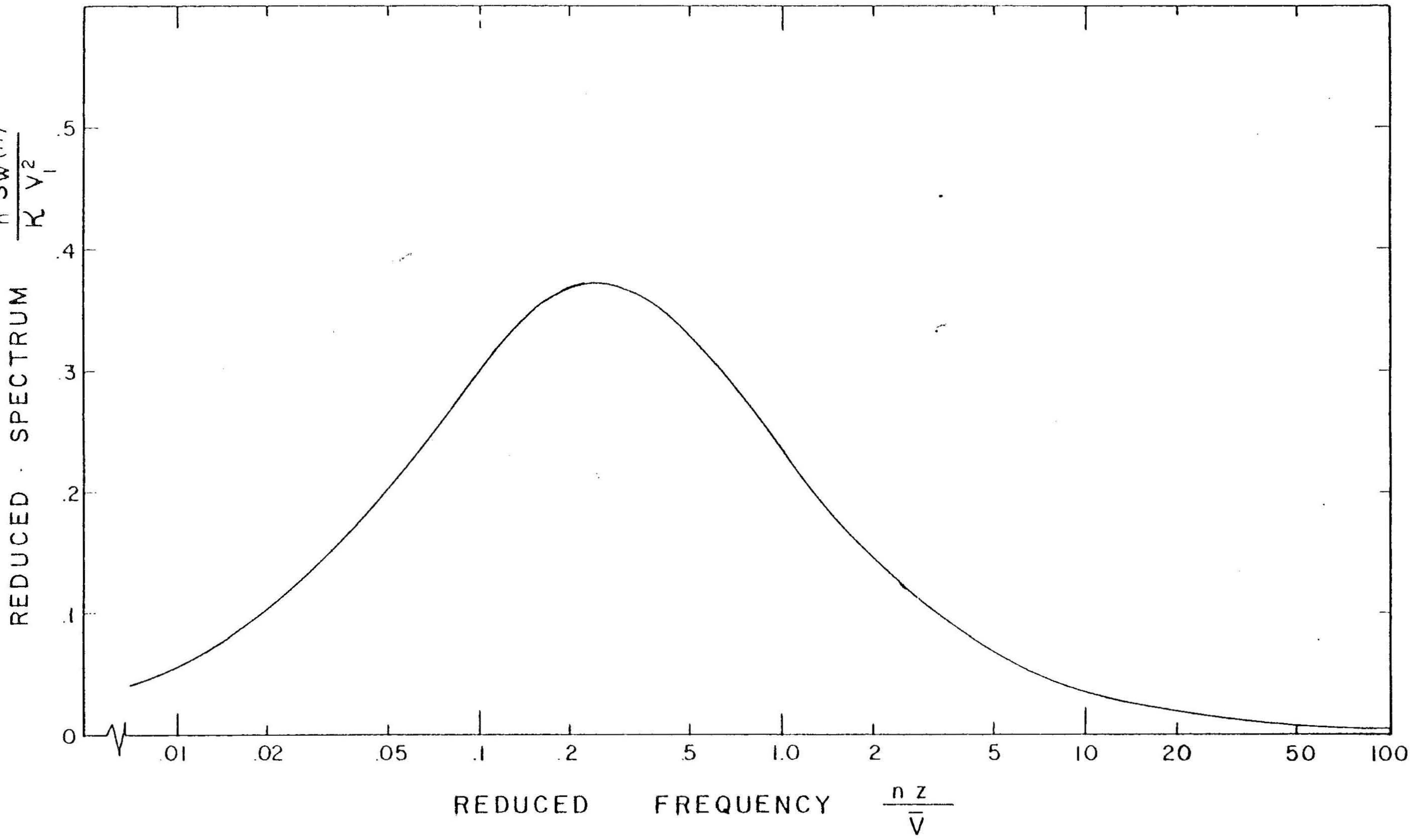


Fig.13. Spectrum of vertical gustiness (after Panofsky).

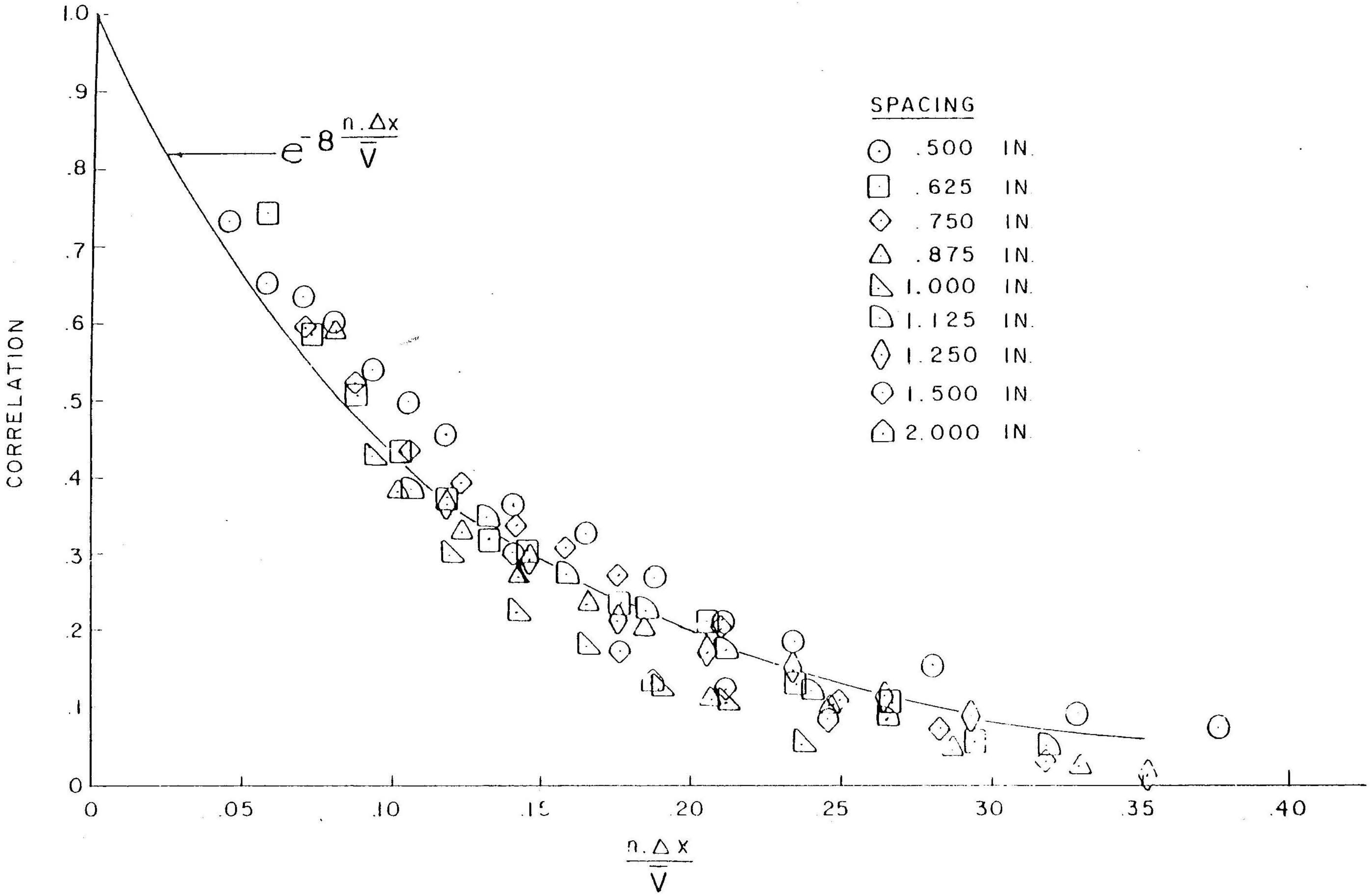


Fig.14. Lateral correlation of longitudinal velocity in strong turbulence in wind tunnel.

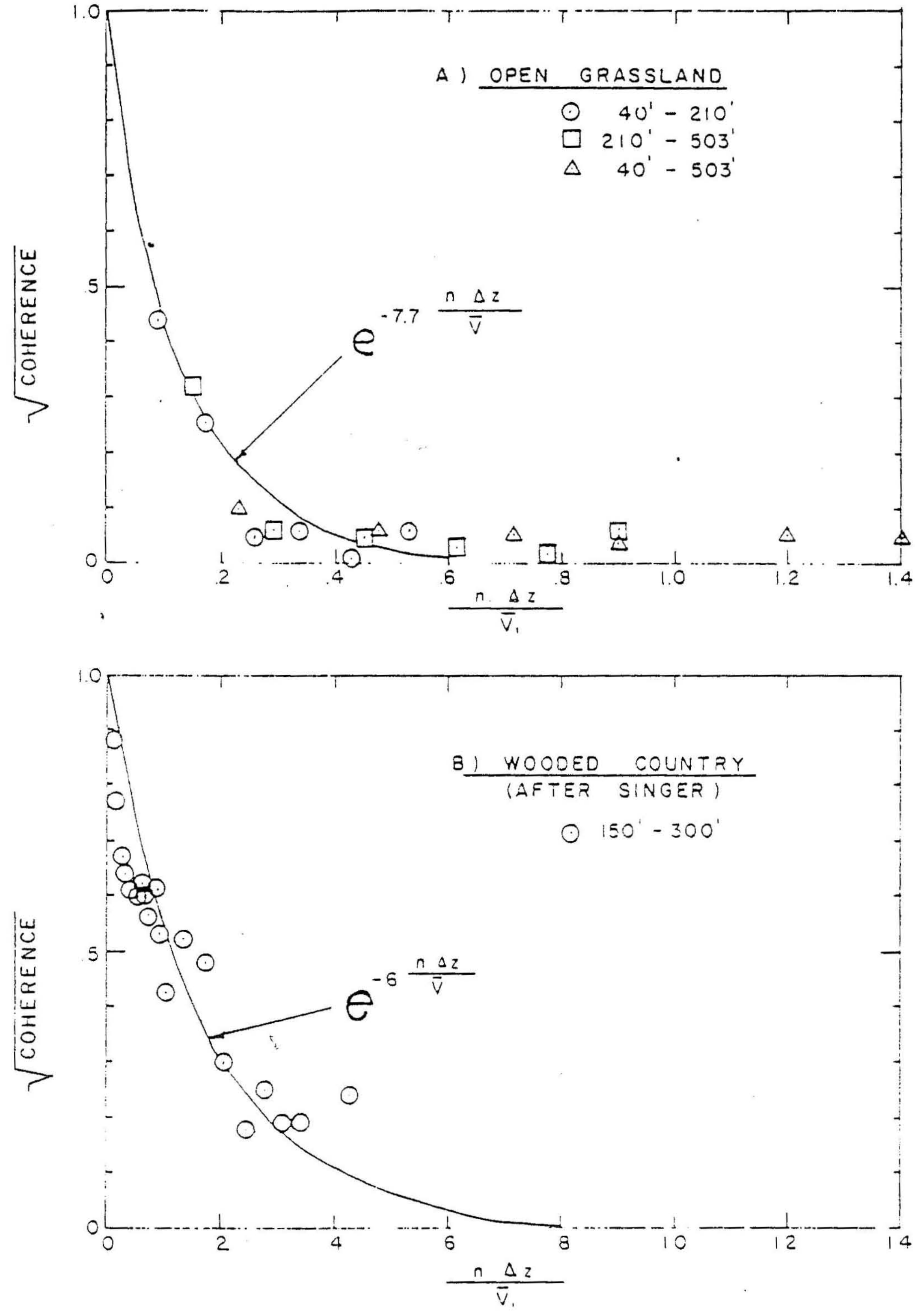


Fig.15. Absolute value of correlation (coherence) of wind speed in vertical direction as function of separation to wavelength ratio.

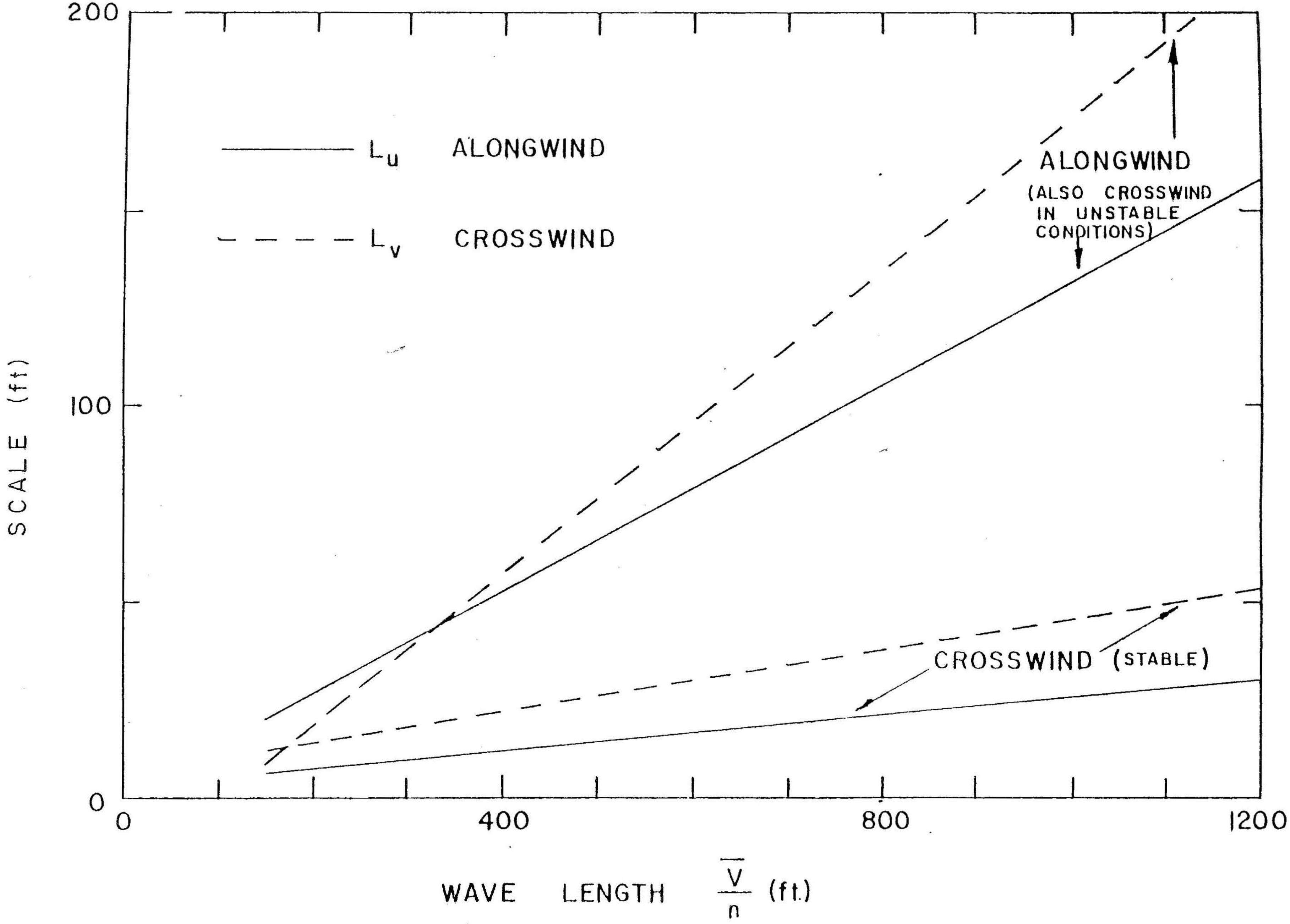


Fig.16. Along-wind and cross-wind scales of turbulence for the U- and V- components of wind velocity as functions of inverse wave number. (After Cramer).

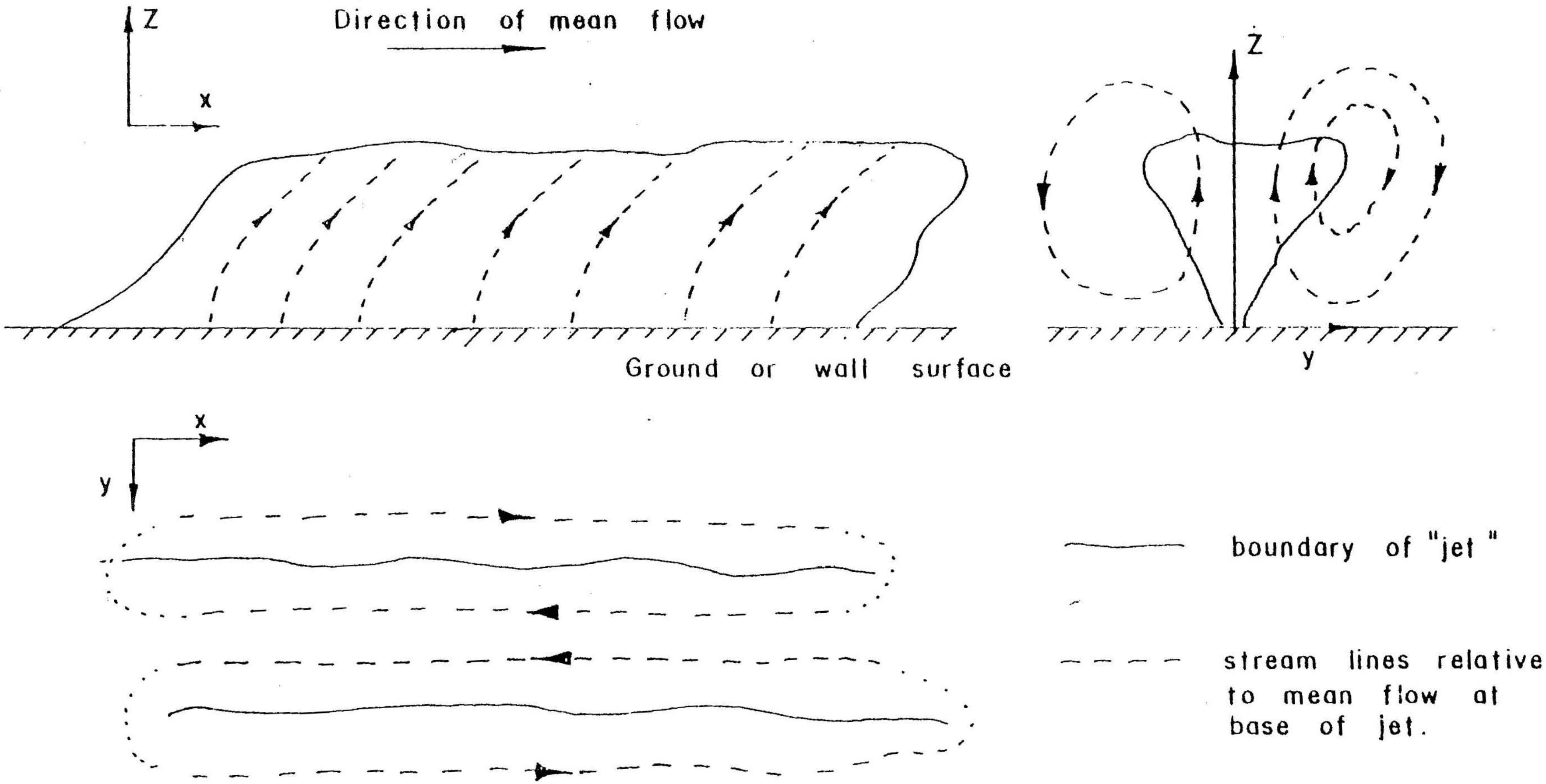


Fig.17. Structure of the "two-dimensional" jets in the constant stress layer (after Townsend - 1957).

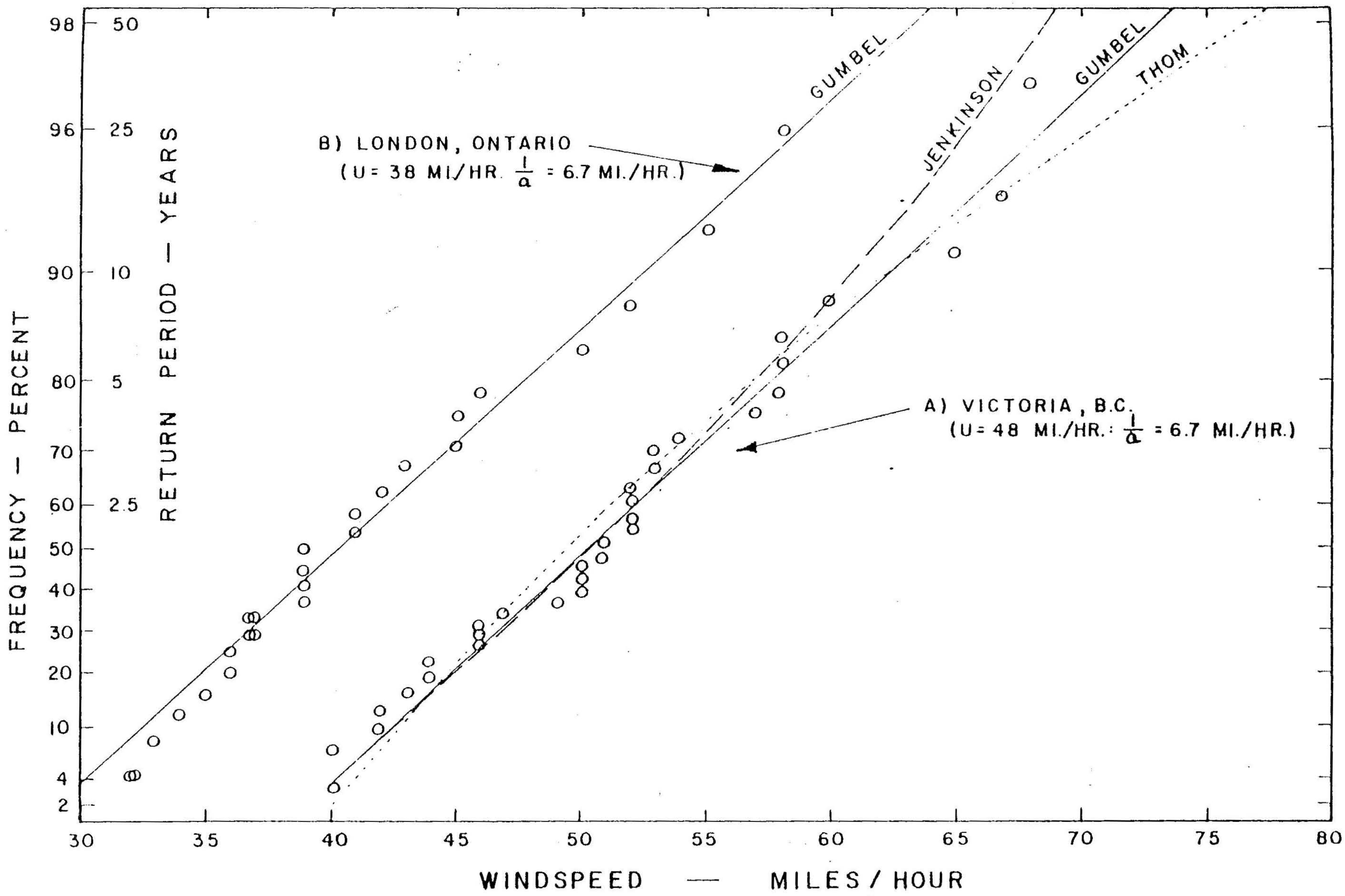


Fig.18. Comparison of annual maximum hourly wind speeds at Victoria B.C. and London, Ontario with theoretical distributions of extreme values (after Boyd and Kendall).

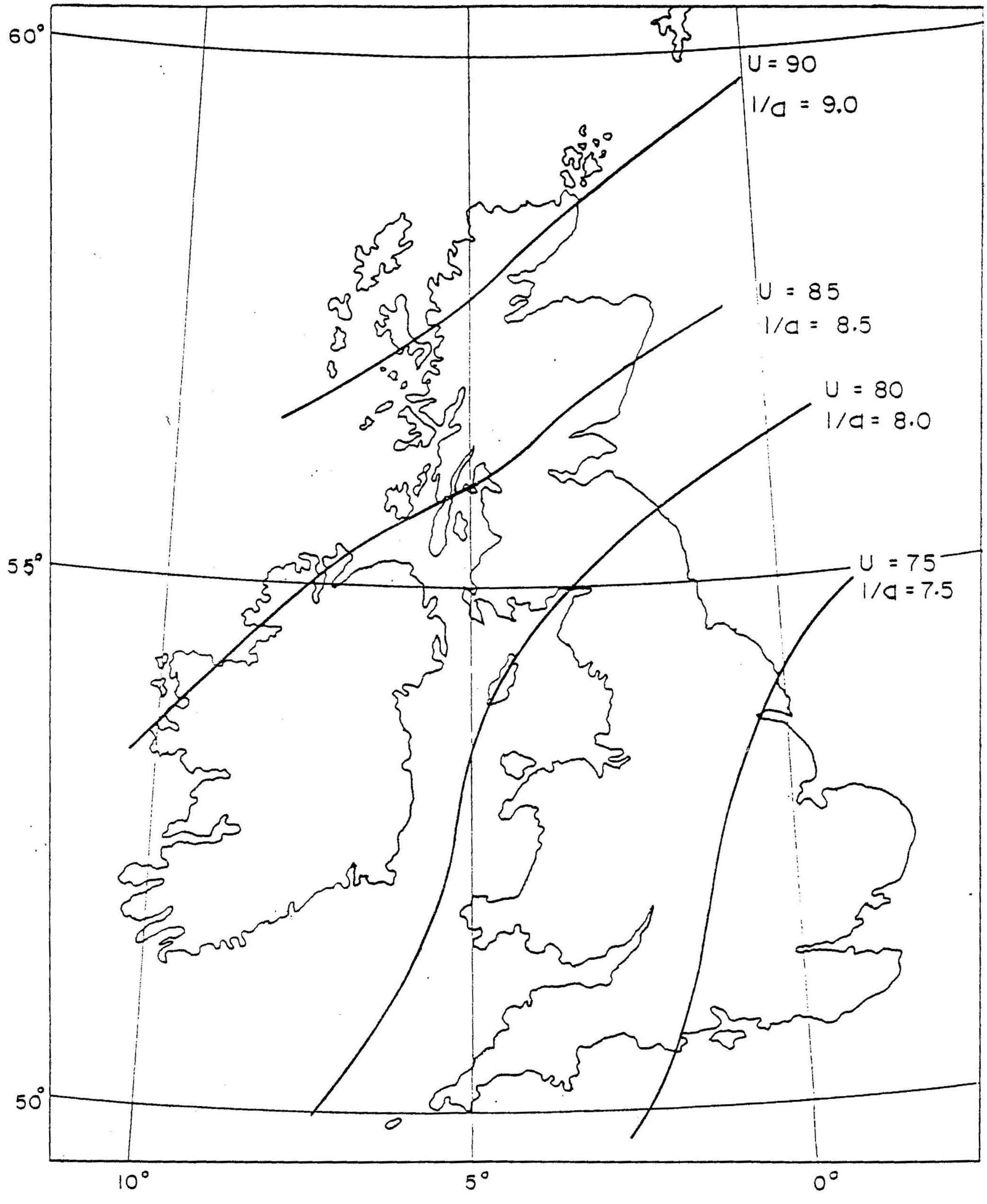


Fig.19. Parameters of extreme mean hourly gradient wind speed over the British Isles.

Units: Miles per hour