Indian J. Met. Hydrol. Geophys. (1975), **26**, 1, 1-10

551. 551

Characteristics of natural wind — Pt. I : Mean flow

V. Kr. SHARAN

Department of Aeronautics,

Royal Institute of Technology, Stockholm

(*Received 27 December 1973*)

ABSTRACT. The aerodynamic friction offered by various objects on the earth and the earth itself to the movement of air creates wind forces on structures. These wind forces have become more and more critical due to the more slender and unusual design of present day structures. The available data on mean wind speed measurement in the lower part of the atmospheric boundary layer, as relevant to wind loading on structures, are analysed critically and correlated analytically. It is concluded that under certain conditions the variation of mean wind speed with height in neutral stability can be satisfactorily represented by a simple power law. The index of the power law and the corresponding gradient height must, however, be varied to take account of the roughness of the surface. Further, the total change in mean wind direction over a few hundred feet being a few degrees, may be neglected during considerations of most structures. Finally suggestions are made for the power law exponent, gradient height and the surface drag coefficient for the four typical categories of terrain. The suggested values are confirmed by recent measurements of wind speed made in various parts of the world.

1. Introduction

Air, chemically speaking, consists of a heterogeneous mixture of nitrogen and oxygen besides small amounts of other gases and water vapour and as such is harmless for civil engineering structures. However, it is continually acted upon by forces of gravitational nature, those generated by heat of the sun, by deflective forces due to earth's rotation, and by centrifugal forces due to curvature of the wind path. The motion of air as a resultant of the above forces is called wind and possesses kinetic energy by virtue of the velocity and its mass. If an obstacle is placed in the path of the wind so that the moving air is stopped or is deflected from its path, then all or part of the kinetic energy of each filament of moving air is transformed into the potential energy of pressure. This results into two kinds of foces acting on the structure, namely : steady wind pressure and dynamic fluctuating pressure. Either or both of them may lead to failure of a structure.

Earth's surface, over which the wind blows, acts as a source of resistance and the flow velocity is reduced in the lower layers dropping to zero adjacent to the surface. This creates varying wind velocity in the vertical direction similar to that of a boundary layer, attaining a constant velocity at gradient height (about 3000 ft according to recent measurements). Besides the resistance offered by earth's surface various structures also offer resistance to the flow of wind. However, these structures vary both in magnitude and in height, thereby offering varying amount of resistance. Because of this and certain other factors air never flows with a perfectly smooth and streamline motion, but always with fluctuations which, when sudden and relatively brief, are called gusts. The masses of air involved in gustiness may simultaneously and in same area cover a wide range of sizes from very small to very large. These fluctuations have vertical as well as horizontal components. It is these aspects of natural wind and their relationship to structures which are discussed in greater detail below.

2. Randomness of natural wind

It is a matter of common experience that natural wind is quite unlike the airflow in a conventional wind tunnel because it is not steady. The natural wind is usually turbulent, consisting of a succession of gusts and lulls with associated fluctuations in wind direction. These changes in speed and direction are very irregular and are made up of oscillations with periods ranging from a fraction of a second to many minutes, the eddies with which they are associated ranging in size from very small to very large.

Wind velocity (speed and direction) vary both with time and space and in space it varies in all three directions. In statistical sense natural wind characteristics provide the best random sample of any phenomenon. Quantitatively this can be deduced from an examination of the traces from a recording anemometer if the time scale of chart is sufficiently open (Fig. 1). This trace is very similar to a record of random electrical noise arising from an electronic apparatus. Two relevant facts are obvious from this chart. The first is that the variations in wind speed which occur have a wide range of time scales ranging from several hours down to fractions of a second. The second is that variations are not regular but are highly complex and irregular, in fact random. In Fig. 1 wind speed varies from 25 to 117 m.p.h. within a few minutes and direction by 160° within a few hours.

This randomness of the variation of wind speed has several important implications, the most obvious being that it is possible to discuss the occurrence of a given wind speed in terms of probability, *i.e.*, a given wind speed has a certain chance of occurring within a specified period of years.

3. Atmospheric stability and strong winds

The static stability of the atmosphere profoundly affects the structure of the wind in the natural boundary layer. However, wind forces acting on structures are significantly large only during strong winds and these occur only during storms.

In strong winds [*e.g.*, U. S. Weath. Bur. Nat. Hurricane Res. Project Rep. No. 5, March 1957; *J. Struct. Div. Proc.* ASCE, **ST4** No. 1708, 1958; *Atmospheric Turbulence*, Methuen and Co. Ltd., 1949; *J. Met.*, **10**(1953), pp. 121–124; *The climate near the ground*, Harvard Univ. Press, 1950; *Bull. Amer. Met. Soc.*, **38**(1957), 6, p. 335; *Quart. J. R. Met. Soc.*, **58**(1932), pp. 285–288] of long duration, in which turbulence causes thorough mixing, the lapse rate near the ground is invariably close to adiabatic (Devenport 1960) corresponding to a state of neutral stability, the exception being severe local storms, such as thunderstorms and frontal squalls. The dominating influence on the velocity profile in these storms is not the stability but the surface roughness. In general, the fluctuations in the flow arise both from mechanical stirring of the mean flow by surface friction and from convection currents caused by thermal gradients in the atmosphere.

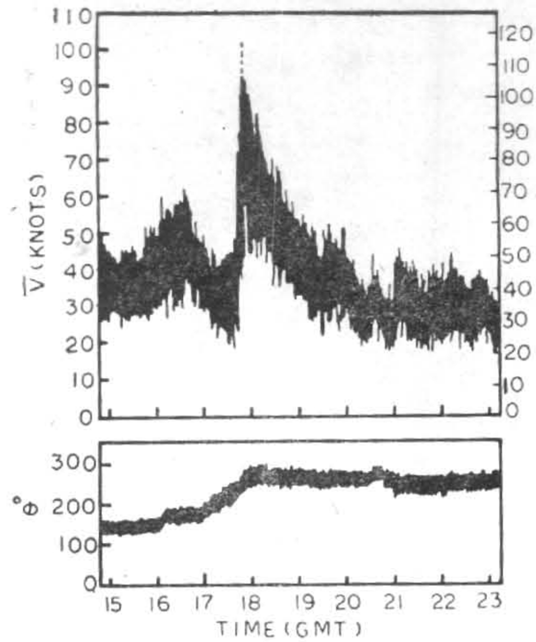

Fig. 1. Record of wind speed and direction at Tiree from 1500 to 2300 GMT, 26 February 1961

However, most structural wind loading problems are concerned with high wind conditions. In high winds, surface friction causes so much mechanical stirring of the atmosphere that the thermal gradients giving rise to convection processes are destroyed, and the lapse rate is always approximately adiabatic and the stability neutral. Under these conditions, the fluctuations in the flow are therefore almost wholly caused by the mechanical stirring of the mean flow by surface friction. From the above it is obvious that for wind loading on structures only the state of neutral atmosphere need be of concern to us.

4. The atmospheric boundary layer

The atmospheric boundary layer arises because of the aerodynamic friction arising from the motion of the air relative to the earth's surface. Above the layer of frictional influence near the surface air moves purely under the influence of pressure gradients and attains what is known as the gradient velocity, also called geostrophic wind. The boundary layer may be regarded as the layer from which momentum is directly extracted and transferred downward to overcome surface friction.

In high winds the boundary layer is fully turbulent, the Reynolds stresses far exceed the direct viscous stresses and the latter may be neglected. Viscosity does, however, play an important role in controlling the rate of dissipation of the turbulence. The rougher the ground, the greater the surface drag force, the turbulent intensity, the Reynolds stresses, the gradient height and the retardation at the surface. Thus the depth of the boundary layer may be regarded as highly variable because the turbulent motion on which the effective transport of momentum depends is controlled both in intensity and in effective vertical range of action by the nature of the surface roughness and the thermal stratification of the atmosphere. The consequence is that by day over land the effective top of the boundary layer is defined by the existence of a layer with stable density stratification beginning at a height in the range of 1500 to 6000 ft. On a clear night with light winds, however, the stable stratification generated by ground cooling may be effective in suppressing the turbulent motion except perhaps immediately above rough ground, the effective boundary layer then being very shallow.

The overall region of frictional layer, called the planetary boundary layer, may be divided into three regions; the 'surface layer' upto about 100 ft, the 'lower layer' upto about 600 ft, and the 'planetary boundary layer' upto about 3000 ft.

In the surface layer, the horizontal stresses can be regarded as constant with height; in the lower layer the wind direction is nearly constant with height, except in very stable air; and in planetary layer above 600 ft the wind speed changes little with height but the wind vector usually turns clockwise with height in the northern hemisphere.

The forces controlling the horizontal mean motion of the atmosphere are the horizontal pressure gradient, the apparent deviating force arising from reference to axes fixed on a rotating earth and the horizontal shearing stresses associated with the vertical transfer of momentum to the ground beneath. The relevant Navier-Stocks equations of motions are,

$$\frac{d\bar{u}}{dt} - f\bar{v} = -\frac{1}{\rho}\frac{\partial\bar{p}}{\partial x} + \frac{1}{\rho}\frac{\partial\tau_x}{\partial z} \qquad (1)$$

$$\frac{d\bar{v}}{dt} + f\bar{u} = -\frac{1}{\rho}\frac{\partial\bar{p}}{\partial y} + \frac{1}{\rho}\frac{\partial\tau_y}{\partial z} \qquad (2)$$

where f is coriolis parameter, $\bar{u}$, $\bar{v}$ are mean wind speeds and τ_x and τ_y are horizontal shear stresses in x and y directions. ρ is density of air, p is mean value of atmospheric pressure.

Theoretical solution of the above equations will depend in detail upon the assumption made about the exchange of momentum. However, it is more fruitful to consider the two extreme cases first. Considering homogeneous steady flow, in which case the first term in each of the above equations is dropped, we get the geostrophic components of the wind, *i.e.*, the velocity components at the top of the boundary layer where the last term in each of the equations is zero (by definition),

$$v_g = \frac{1}{\rho f}\frac{\partial\bar{p}}{\partial x} \qquad (3)$$

$$u_g = -\frac{1}{\rho f}\frac{\partial\bar{p}}{\partial y} \qquad (4)$$

where u_g and v_g are geostrophic wind components.

The resultant V_g directed along the isobars, is customarily regarded as an approximation to the external of free wind unaffected by surface friction.

For small values of z, assuming that $\partial\bar{p}/\partial x$ is independent of z and that the turning of wind direction with height is small, we get, by integrating equation (1) the change in resultant stress over the height range 0-z,

$$\frac{\tau(0) - \tau(z)}{\tau(0)} = \frac{\rho f z\, V_g \sin\alpha_g}{\tau(0)} \qquad (5)$$

where α_g is the total turning of wind in the boundary layer. From observations (Pasquill 1970) $\sin\alpha_g = 0\cdot3$ and $\tau_o/\rho V_g^2 \approx 10^{-3}$, we get for $V_g = 33\cdot3$ ft/s, and $f \simeq 10^{-4}$ for middle latitudes,

$$\frac{\tau(0) - \tau(z)}{\tau(0)} = 10^{-3}.\, z \qquad (6)$$

with z in ft. Thus we have a fall in stress away from the boundary which remains negligible, say less than 10 per cent, for heights upto about 100 ft.

This defines the so called constant stress layer (earlier referred to as the surface layer), a condition which has important consequences in the theoretical analysis of low-level wind structure. Sheppard (1970) also defined an interfacial layer as the layer within the roughness elements in which downward flux of momentum is handed over to pressure forces on the elements themselves, a layer of special interest here.

We will now study the details of the wind structure in the boundary layer above the interfacial layer, both theoretically and experimentally.

5. The variation of mean wind velocity with height

In the surface layer, based on the assumption of constant shear stress and that the mixing length is proportional to the height, in conditions of neutral stability, we may write the wind velocity profile as,

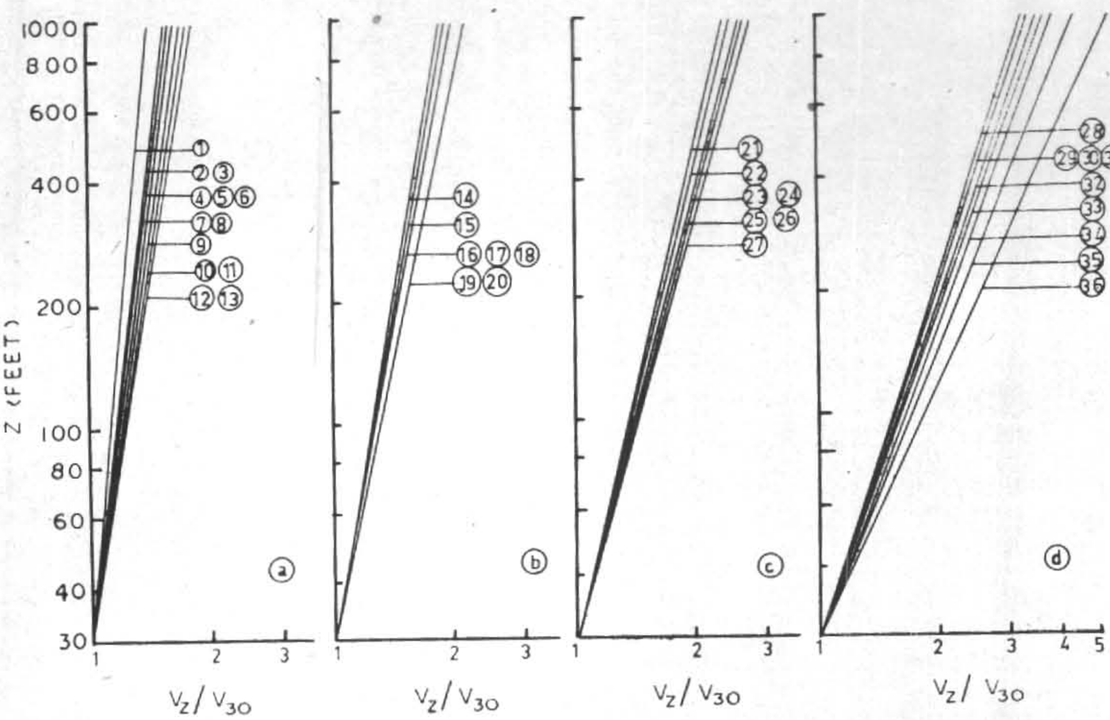

Fig. 2(a-d). Effect of surface roughness on rate of increase of mean wind velocity with height. For other details *see* Table 1

$$\frac{V_{z1}}{V_{z2}} = \text{Const} \times \log\left(\frac{z_1}{z_2}\right) \tag{7}$$

where V_{z1} and V_{z2} are mean wind speeds at heights z_1 and z_2.

Several approaches have been taken to formulate the mean wind profile through the deeper part of the natural boundary layer, using various concepts of eddy viscosity. An assumption of constant eddy viscosity has led to the so called Ekman spiral. Other methods using power law variations of eddy viscosity with height have been put forward by Pranditl, and Tolmien and Kohler (*see* Sutton 1949, 1953). A two-layer model which implies a mixing length increasing linearly with height to the top of the surface layer and above that decreasing linearly to the top of the planetary layer, has been suggested by Ross and Montgomery (Ross and Montgomery 1935). However, for neutral stability conditions the resulting complexities do not appear to be justified, since it is possible to obtain a good fit to experimental data by means of simple power law of the form,

$$\frac{V_{z1}}{V_{z2}} = \left(\frac{z_1}{z_2}\right)^{\alpha} \tag{8}$$

where α is power law exponent.

In principle, a law of this type relates the mean wind velocities at any pair of heights z_1 and z_2. In practice, it is useful to standardize the presentation by using either the gradient height or a standard height of 30 ft as a reference. In these cases the power law becomes,

$$\frac{V_z}{V_g} = \left(\frac{z}{z_g}\right)^{\alpha} \tag{9}$$

and

$$\frac{V_z}{V_{30}} = \left(\frac{z}{30}\right)^{\alpha} \tag{10}$$

where V_g is resultant geostrophic wind speed and V_{30} is mean wind speed at 30 ft.

At present time Eq. (9) is not so useful in practice, because of uncertainty in the determination of the gradient height z_g.

Fig. 2(a,b,c,d) shows the mean velocity profile measurements from 36 different workers over different terrains. The profiles of special importance in the design of tall buildings are those for city centres plotted in Fig. 2(d). The corresponding exponents are shown in Table 1 (a,b,c,d). It should be noted that the records refer specifically to mean velocity profiles prevailing above a height of 30 ft in strong winds, over flat ground surface in lapse rate believed fairly close to adiabatic. The records, therefore, are homogenous except that the nature of the ground roughness

TABLE 1

Influence of surface roughness on power law exponent

Curve No.	Location	Power law exponent, α	Reference
	(a) Flat open country and open water		
1	Caspian Sea	0·10	Goptarev (1957)
2	Ballybunion, Ireland	0·11	Wing (1921)
3	Masned Sund, Denmark	0·11	Jaul (1954)
4	Salisbury Plain, U.K.	0·13	Scrase (1930)
5	Cardington, U.K.	0·13	Giblett *et al.* (1932)
6	Dallas, U.S.A.	0·13	Davenport (1967)
7	Ann Arbor, U.S.A.	0·14	Sherlock (1953)
8	Salisbury Plain, U.K.	0·14	Pagon (1935)
9	Cardington, U.K.	0·15	Sutton (1949)
10	Suffield, Alta.	0·16	Davenport (1967)
11	Sale, Australia	0·16	Decon (1955)
12	Leafield, U.K.	0·17	Heywood (1931)
13	Savannah River, U.S.A.	0.17	Ligtenberg (1961)
	(b) Treed and rough country		
14	Lopik, Holland	0·18	Decon (1955)
15	Honshu, Japan,	0·19	Decon (1955)
16	London, Ontario	0·20	Davenport (1967)
17	Clovis, N.M.	0.20	Davenport (1967)
18	Orkney Island, U.K.	0·20	Wax (1956)
19	Quickborn, Germany	0·22	Franckenberger and Rudloff (1949)
20	Akron, U.S.A.	0·22	Huss and Portman (1949)
	(c) Large towns and small cities		
21	Upton, L.I., U.S.A.	0·25	Smith (1953)
22	Brookhaven, U.S.A.	0·26	Davenport (1967)
23	Montreal, Canada	0·28	Davenport (1967)
24	St. Louis, U.S.A.	0·28	Davenport (1967)
25	Tokyo, Japan	0·29	Davenport (1967)
26	Brookhaven, U.S.A.	0·29	Davenport (1967)
27	Upton, L.I., U.S.A.	0·30	U.S. Weather Bur. (1957)
	(d) Centres of large cities		
28	Tokyo, Japan	0·34	Kawanabe (1964)
29	Brookhaven, U.S.A.	0·35	Davenport (1967)
30	Kiev, USSR	0·35	Shitoni and Yamamoto
31	Farnborough, U.K.	0·35	Dines (1912-1913)
32	Moscow, USSR	0·37	Ivanov and Klinoov (1961)
33	Copenhagen, Denmark	0·38	Jensen (1958)
34	New York City, U.S.A.	0·39	Rathburn (1940)
35	Leningrad, USSR	0·41	Ariel and Kliuchnikova (1960)
36	Paris, France	0·45	Pagon (1935)

and the aggregate nature of obstructions vary widely, from the smooth surface of the sea to the rough obstructed surface of a large city.

The exponent of the power law increase is seen to vary between 0·10 and 0·41 [the only exception being Paris, for which $\alpha = 0{\cdot}45$; here the measurements are not believed to be very reliable being from Eiffel (1900) when the instrumentation for wind speed measurements was not accurate], depending only on the surface roughness characteristics. It is seen that curves 1 to 13 (Fig. 2 a) have exponents lying between 0·10 and 0·17 and, in each case, the surrounding terrain was characteristically flat and open. The average value for the above observations taken on land lies close to 0·14. The exponent of curves 14 to 20 (Fig. 2 b) lies between 0·18 and 0·22 with an average of 0·20. For these curves the terrain corresponds to rougher characteristics of treed and wooded farmland, towns, scrub trees, city outskirts, etc. The exponent of curves 21 to 27 (Fig. 2 c) lies between 0·25 and 0·30 with an average of 0·28. These curves are derived from records of large towns and small cities. The exponent of curves 28 to 35 lies between 0·34 and 0·41 with an average of 0·37. These curves are based on records of large cities (Table 1 d) which represents conditions of extreme surface roughness. For all the curves the value of the exponent is seen to increase in a consistent manner with the increase in the roughness of the terrain, thereby illustrating the effect of surface roughness on mean velocity profile.

The value of surface friction is known to increase slightly with velocity and this may cause a slight secondary effect on the mean profile shape. Collins (1955) found that the exponent of the power law profile increased by approximately 0·02 for every 10 mph increase in surface wind velocity; at 50 mph the value being 0·27 and the extrapolated value at 80 mph being 0·33. These results also indicate that the effect of wind velocity is only of secondary importance as compared to that of surface roughness.

Based on the above results the following values for the power law exponent for different terrains are suggested (Table 2). Compare these values of α and z_g with those generally accepted (Table 3).

These values refer to the mean wind velocity over level ground, to large scale severe storms exhibiting nearly neutral stability and to heights between about 30 ft and the height at which the gradient velocity is first attained. This height, also called the gradient height, may be

TABLE 2

Average parameters of power law profile and surface drag for various type of terrains

Type of terrain	Power law-exponent α	Gradi-ient height z_g (ft)	Surface drag coefficient
1. Open terrain with very few obstacles, *e.g.*, open grass or farmland with few trees, hedgerows and other barriers etc. prairie, tundra shores and low island of inland lakes etc—called terrain A	0·14	900	0·003-0·005
2. Terrain uniformly covered with obstacles 30-50 ft high *e.g.*, residential, small towns wooded shrub, small fields with bushes, trees and hedges — called terrain B	0·20	1300	0·015-0·025
3. Terrain covered with large regular objects such as found in large towns and small cities consisting of low roofed houses and low-rise buildings — called terrain C	0·28	1500	0·025-0·035
4. Terrain with large irregular objects, *e.g.*, centers of large cities, very broken country with many wind breaks of tall trees etc — called terrain D	0·37	1700	0·035-0·050

TABLE 3

Commonly accepted values of power law profiles etc for various terrains

Type of terrain	α	z_g (ft)	K
A	0·16	900	0·005
B	—	—	—
C	0·28	1300	0·020
D	0·40	1700	0·050

TABLE 4

Observed ratios for surface to gradient wind

City	Observed ht. (ft)	Observed exponent	$\frac{V}{V}$obs	Reference
Washington	100	--	0·45	Graham and Hudson (1960)
Kiev	66	0·34	0·40	Ariel and Kliuchnikova (1960)
Brookhaven	355	0·32	0·61	Davenport (1967)
Leningrad	492	0·41	0·70	Ariel and Kliuchnikova (1960)

determined from the ratio of the surface to gradient velocity. Observed values of these ratios for city centres (terrain-D) are given in Table 4 and plotted in Fig. 3(a). The results fall on a line having an exponent of 0·36 and intercepting the gradient velocity at 1700 ft. This value of α is in agreement with the value suggested in Table 2. Other generally accepted values of gradient height for terrains A, B and C are also shown in Table 2, along with the surface drag coefficient K for various types of surface. The surface drag coefficient is usually defined from the relation between the shear stress near the ground, and the mean wind speed. It may be regarded as a dimensionless measure of the roughness in the same way that the conventional aerodynamic drag coefficient is a measure of the drag of a body.

Most of the available wind speed data is from measurements carried out at airports located well outside the city, where the terrain corresponds to category-A since it, generally, is a flat open terrain with very few obstacles. Table 5 shows data on extreme wind speeds (Davenport 1967) observed at airports (Terrain-A) and city centres (Terrain-D). The effect of increased surface roughness in the city on wind speed is obvious (the wind speed being lower in the city in spite of the anemometer height being higher in the city than at the airport). The values from the airport fit very closely the power law profile ($\alpha=$ 0·14) with z_g=900 ft (Fig. 3 b), thereby confirming the values for terrain-A given in Table 2. Using the gradient wind speed from the above, the city centre wind speeds were normalized and are shown in Fig. 3 (b). They fit very closely the power law curve with an exponent of 0·36 and z_g=1700 ft; compare with the values given in Table 2. The above measurements confirm the values suggested in Table 2. The mean velocity profile curves corresponding to the above four types of terrain along with their respective exponents and gradient heights are shown in Fig. 4.

We may summarize the discussion as follows. Provided that the terrain is reasonably level, and of sufficiently uniform surface roughness to allow a state of dynamical equilibrium to be established between the drag and stirring action of the surface, and the steady flow at high level determined by the isobar map, the variation of mean wind speed with height in

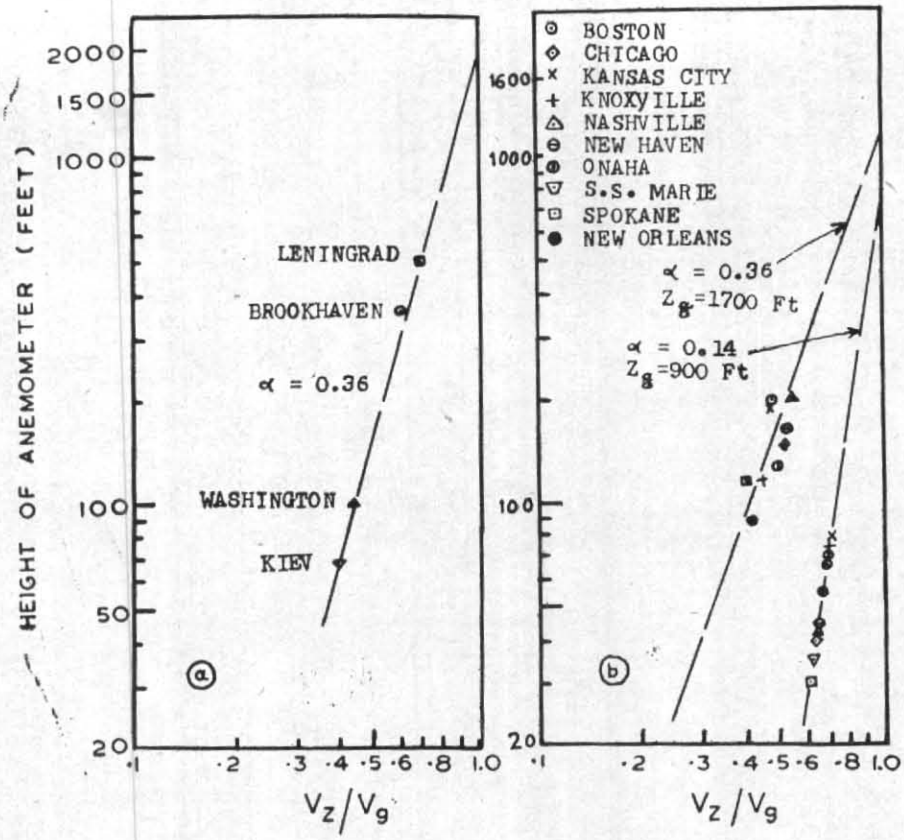

Fig. 3 (a). Observed ratios of surface to gradient wind speed over rough terrain

Fig. 3 (b). Comparison of fastest mile wind speed at city and airport stations for US cities

neutral stability conditions can be satisfactorily determined by a simple power law. The index of the power law and the corresponding gradient height must, however, be varied to take account of the roughness of the surface. However, two problems associated with the local topography (since the obstructions concerned are several gradient heights in extent) should also be taken into account. They are : change in the mean wind speed characteristics associated with a sudden change of roughness (*e.g.*, when the wind blowing over open country suddenly encounters the suburbes of a city), and the change in the wind structure produced by presence of hills (Taylor 1962, Panofsky 1964).

Fig. 5 shows detailed measurements of mean wind speed over London (Newberry 1971, Hilliwell 1968), Hong Kong, Leipzig (Lettau 1950) and Rugby (Harris 1970). Wind data over London corresponds to terrain-D and that over Hong Kong to terrain-A data of Leipzig corresponds to that of terrain-B and that over Rugby to terrain type between A and B. The corresponding values of α shown in Fig. 5 are in agreement with the values suggested in Table 2.

6. Hourly mean wind direction

The solution of the equation of motion of a fluid on a rotating earth predicts a systematic change in wind direction, clockwise in the northern hemisphere, between ground level and the gradient height. The amount of rotation predicted is dependent upon the assumptions made about the exchange of momentum due to turbulence at various heights above ground level. An earlier theory by Ekman using the simplest assumption of constant eddy viscosity predicted a rotation of 45° between ground level and z_g. More sophisticated theories (*e.g.*, Sutton 1953) are also available, which take into account the influence of the nature of the terrain. Using a power law variation of eddy viscosity with height, Sutton (1953) obtained the following relation for the angle between the surface wind and the geostrophic wind for various type of terrains,

$$\tan\ \theta^\circ = \sqrt{\alpha\ (1+\alpha)} \tag{11}$$

where θ is change in wind direction.

Fig. 6 shows mean values of wind direction from pilot balloon data (Lettau 1950) over Leipzig,

TABLE 5

Once-in-50-year wind speeds in U.S. cities

City	Airports		City Office	
	Anemometer ht. (ft)	Wind speed (mph)	Anemometer ht. (ft)	Wind speed (mph)
Boston	63	103	188	72
New Haven	42	74	155	60
Chicago	38	70	—	57
S.S. Marie	33	85	52	63
Kansas City	76	95	181	63
Omaha	68	91	121	65
Knoxville	71	89	111	57
Nashville	42	86	191	73
Spokane	29	78	110	51

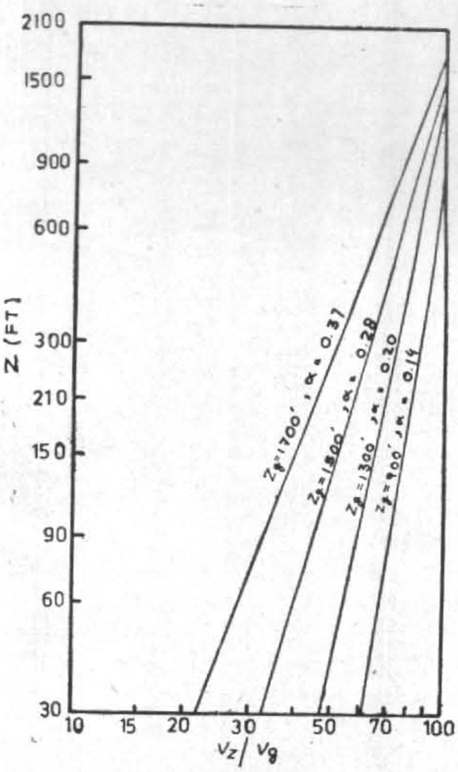

Fig. 4. Increase of wind speed with height for different types of surface roughness according to Eq. (9)

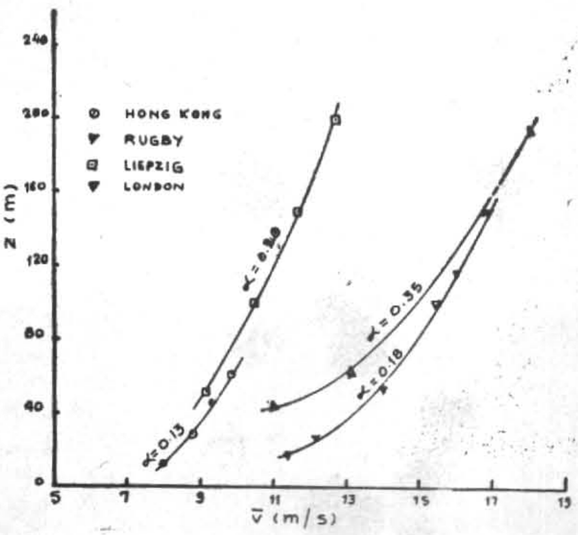

Fig. 5. Measured wind profiles over Hong Kong, Rugby, Leipzig and London

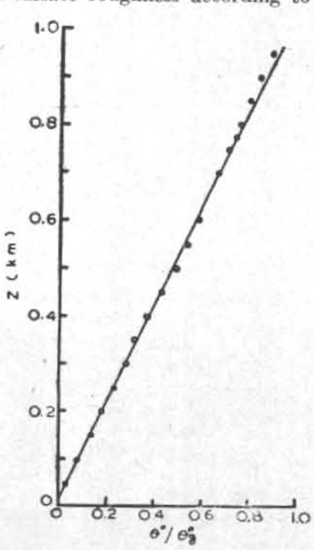

Fig. 6. Variation of mean wind direction with height over Leipzig

expressed as the ratio between the surface and the geostrophic directions. Typical feature is the relatively uniform gradient of direction over the whole depth of the boundary layer. However, it should be noted that over 800 ft, the tallest structure likely to be erected, total change in the wind direction is about 5 to 6°, and hence may be neglected during considerations of most buildings. This fact is confirmed by detailed measurements of wind direction over Rugby Radio Station by Harris (1970). Fig. 7 shows values of mean wind direction measured at Rugby Radio Station in neutral stability condition. It is obvious that there is little change of direction with height from ground level to a height of 590 ft. This observation has been confirmed by other similar results also. However, except for the design of very tall buildings, or

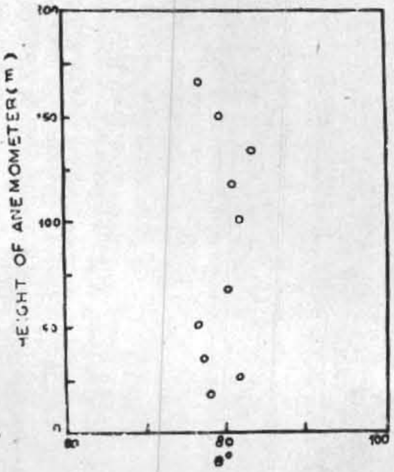

Fig. 7. Measured mean wind direction at Rugby

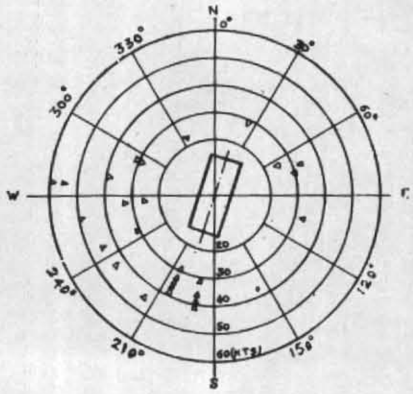

Fig. 8. Polar diagram of winds measured at Royex house in London

those with special features, making them very sensitive to wind direction, it seems reasonable to ignore the change of wind direction with height for high wind conditions and all types of terrain. On the other hand, the wind direction and wind speed both change continuously with time (Newberry 1968), when measured at a point (Fig. 8). This shows that although the wind direction may be assumed constant with height for all practical purposes, there is no well defined wind direction for a given site at a given time. This fact is of great significance in the statistical treatment of wind data.

7. Conclusions

From the above, one may conclude the following:

The total wind speed vary both with time and space and the variations are random in nature. This randomness of natural wind makes it possible to discuss the occurrences of a given wind speed in terms of probability during considerations of buildings and structures.

The large scale mature storms, causing strong winds, exhibit nearly neutral stability and therefore the dominating influence on the velocity profile in high winds is not the stability but the surface roughness.

The overall region of frictional influence may be divided into three regions : the surface layer upto 100 ft, the lower layer upto 600 ft and the planetary boundary layer upto 3000 ft. The variation of mean wind speed with height, in the entire boundary layer, under certain conditions, in neutral stability may be satisfactorily represented by a simple power law. The index of the power law and the corresponding gradient height, however, vary depending on the roughness of the terrain, local topography and wind direction.

The terrain may be classified into four categories : (*a*) flat open country and open water ($\alpha=0\cdot14, z_g=900$ ft), (*b*) treed and rough country ($\alpha=0\cdot20$, $z_g=1300$ ft), (*c*) large towns and small cities ($\alpha=0\cdot28$, $z_g=$ 1500 ft) and (*d*) centres of large cities ($\alpha=0\cdot37$, $z_g=1700$ ft).

The total change in wind direction over a few hundred feet may be neglected during considerations of most structures.

Acknowledgements

Acknowledgements are due to Miss I. Ademan and Mrs. B.M. Constantinides for collecting the references. The data for London measurements (Fig. 5) was kindly supplied by Mr. C.W. Newberry of Building Research Station, Watford, England, and the data for Hong Kong measurements was very kindly supplied by Prof. S. Mackey of the University of Hong Kong.

REFERENCES

Ariel, N. A. and Kliuchnikova, L. A. 1960 'Wind over a city', USSR Glavnia Geofizicheskaia Observatoria in A.I. Valikova, Trudy Vypusk, 94, pp. 29-32.

Collins, G. F. 1955 Determining basic wind loads. Proc. ASCE, **81**.

Davenport, A. G. 1960 J. of Structural Div., Proc. ASCE, **ST5**, pp. 39-68.

1967 'The dependence of wind loads on meteorological parameters' Proc. of Int. Res. Seminar on wind Effects on Buildings and Structures, Ottawa.

REFERENCES (contd)

Decon, E. L. 1955 *Quart. J. R. Met. Soc.*, **81**. pp. 562-573.

Dines, J. J. 1912 1913 4th report on wind structure', ARC, R and M, pp. **310**-330.

Franckenberger, E. and Rudloff, W. 1949 'Wind Messungen an den Quick-borner Funkmasten 1947-8, *Met. Amt. Fur N.W. Deutsch.*, Hamburg.

Giblett, M. A. *et al.* 1932 *Geophys. Mem.*, 54, Met. Office, London.

Goptarev, N. P. 1957 'Some results of the investigation of gradients in the vicinity of oil drilling derricks', Trudy Gosudarst venny, Okeanograficheskу Institut, Moscow, Vypusk, 36, p. 128.

Graham, H. E. and Hudson, G. N. 1960 Nat. Hurricane Res. Project, Rep. 39, U.S. Weath. Bur.

Harris, R. I. 1970 'The nature of the wind', A CIRIA Seminar on 'The modern design of wind-sensitive structures'.

Helliwell, N. 1968 'Some open scale measurements of wind over central London', WMO Symp. Brussels.

Heywood, G. S. P. 1931 *Quart. J.R. Met. Soc.*, **67**, pp. 433-455.

Huss, P. O. and Portman, D. J. 1949 'Daniel Guggenheim Airship Inst. Rep. No. 156', Akron Univ., Ohio.

Ivanov, V. N. and Klinov, F. Ya. 1961 'Some characteristics of a turbulent velocity field in the lower 300 m layer of the atmosphere, *Izv. Geophy. Ser.*, pp. 1570-1577.

Jensen, M. 1958 'The model law for phenomena in natural wind' Ingenioren, International Ed. **2**, 4, pp. 121-128.

Juul, J. 1954 'Measuring wind velocities at varying heights above ground', Final rep. of working group No.2 on 'Wind power, Organization' for European Economic Coop.

Kawanabe, Y. 1964 *Bull. Kobe Mar. Obs.*, **172**, pp. 93-118.

Lettau, H. 1950 *Tellus*, **2**, pp. 125-129.

Ligtenberg, F. K. 1961 'Windtrilligen in hoge Gebouwen, Heron, **9**, pp. 22-43.

Newberry, C. W., Eaton, K. J. and Mayne, J. R. 1968 'The nature of gust loading on a tall building', BRS CP 66/68.

1971 'Wind pressures on the post-office tower, London', BRS CP 37/71.

Panofsky, H. A. and Townsend, A. A. 1964 *Quart. J. R. Met. Soc.*, **9**, pp. 147-155.

Pagon, W. W. 1935 *Engng. News Record*, **114**, 21, pp. 742-751.

Pasquill, F. 1971 *Phil. Trans. Roy. Soc. Lond.*, **A269**, pp. 439-456.

Rathburn, J. C. 1940 'Wind forces in a tall building', Trans. ASCE, pap. No. 2050, 105, pp. 1-41.

Ross, C. G. and Montgomery, R. B. 1935 Papers in Physical Oceanogr. and Met., MIT and Woods Hole Oceanographic Inst., 3, Pt. 3.

Scrase, F. J. 1930 *Geophys. Mem.*, 52, Met. Office, London.

Sheppard, P. A. 1970 'The atmospheric boundary layer in relation to large-scale dynamics in the global circulation of the atmosphere', (Ed. G.A. Corby), R. Met. Soc.

Sherlock, R. H. 1953 'Variation of wind velocity and gusts with height', Trans. ASCE, pap. No. 2553.

Shitoni, M. and Yamamoto, I. — 'Atmospheric, turbulence over the large city', 2nd Report, Japan.

Smith, M. E. 1953 Discussion of 'Variation of wind velocity and gusts with height', *Trans. ASCE*, **118**, p. 493.

Sutton, O. G. 1949 *Atmospheric turbulence*, Methuen and Co. Ltd.

1953 '*Micrometeorology*, McGraw-Hill Publ. Co. Ltd.

Taylor, R. J. 1962 *J. Fluid Mech.*, **13**, p, 529.

Wax, M. P. 1956 'An experimental study of wind structure with reference to the design and operation of wind driven generators', British Electrical and Allied Ind. Res. Assoc., Tech. Rep. C/T114.

Weather Bureau (U.S.) 1957 Nat. Hurricane Res. Project, Rep. 5.

Wing, S. P. 1921 'Wind pressures and the design of radio and high transmission towers, *Electrician*, **87**, No. 2256, pp. 6-10.