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CHAPTER I

INTRODUCTION

The Importance of Wind

Wind is one of the geological processes that acts on the surface of the earth. It is a major factor in the geomorphological formation of soil. Wind is a media of the exchange of mass, heat, and momentum between the surface and the environment. It also affects the ecological system of the earth through the transport of seeds, micro-organisms, pollutants, air circulation and even the weather system.

Winds cause forces which change the microtopography and the topography of the surface of the earth and creates a serious erosion problem in many areas of the world. It removes the fertile top soil of farm lands and deposits sediment on arable lands reducing their availability. Sand laden wind also causes physical damage to plants especially during the critical period following emergence.

Wind transports water vapor away from the canopy boundaries, thereby decreasing vapor density and temperature and increasing the vapor pressure gradient around the transpiring leaves. This can result in lower yield. In windy areas, about 60 percent of the evapotranspiration is caused by the wind while 40 percent is from radiation components.

A strong wind removes heat by conduction. This effect is explained by the wind chill index. A 30 mph wind at a temperature of $10^{\circ}F$ can cause the same heat loss from exposed skin as an equivalent temperature of $-33^{\circ}F$ with no wind (USDA 1969).

Blowing dust is a major source of particulate pollution of the air and decreases visibility. Wind movement over wastewater treatment facilities such as clarifiers and filters causes short circuiting which results in decreased performance of the units. Blowing snow accumulates on highways and results in hazardous driving conditions. A thorough understanding of wind systems is important if effective control methods are to be developed.

The velocity profile of wind near the surface of the earth can be described by the following equation:

$$U = \frac{U_*}{k} \ln \left(\frac{Z - D}{Z_o}\right) \qquad (1.1)$$

where U = average velocity measured at height Z,

U_* = shear velocity,

Z = height of velocity measurement,

D = displacement height, and

Z_o = aerodynamic roughness.

The two parameters D and Z_o must be known in order to accurately predict the shape of the velocity profile. Neither displacement height nor aerodynamic roughness have been defined sufficiently to make accurate predictions of these parameters. Therefore, a need exists to develop accurate prediction equations for these parameters.

Objectives

The objectives of this research are:

1. To physically define the displacement height.
2. To develop a prediction equation for displacement height that is physically based and can be applied to various surface conditions.
3. To develop a prediction equation for aerodynamic roughness.
4. To verify the developed equations using published data.

The General Atmospheric Layer

The atmosphere can be divided into three layers as shown in Figures 1 and 2. The upper layer exists above the top of the boundary layer and is not affected by surface conditions. Air flow below the upper layer is characterized by turbulence.

Turbulence is defined as a mode of fluid motion that is a three-dimensional continuum phenomenon, rotational, dissipative, nonlinear, stochastic, and diffusive with the transport occurring at time and length scales typically the same as those of the properties being transported (Vinnichenko et al., 1973).

The intermediate layer exists between the top of the surface layer and the top of the boundary layer. It is characterized by unstable and convective or heat generated turbulence. The lower layer is in contact with the surface and often dominated by mechanical or surface generated turbulence. This layer, where most human interests exist, is estimated (Hagen, 1980) to be one-tenth of the depth of the

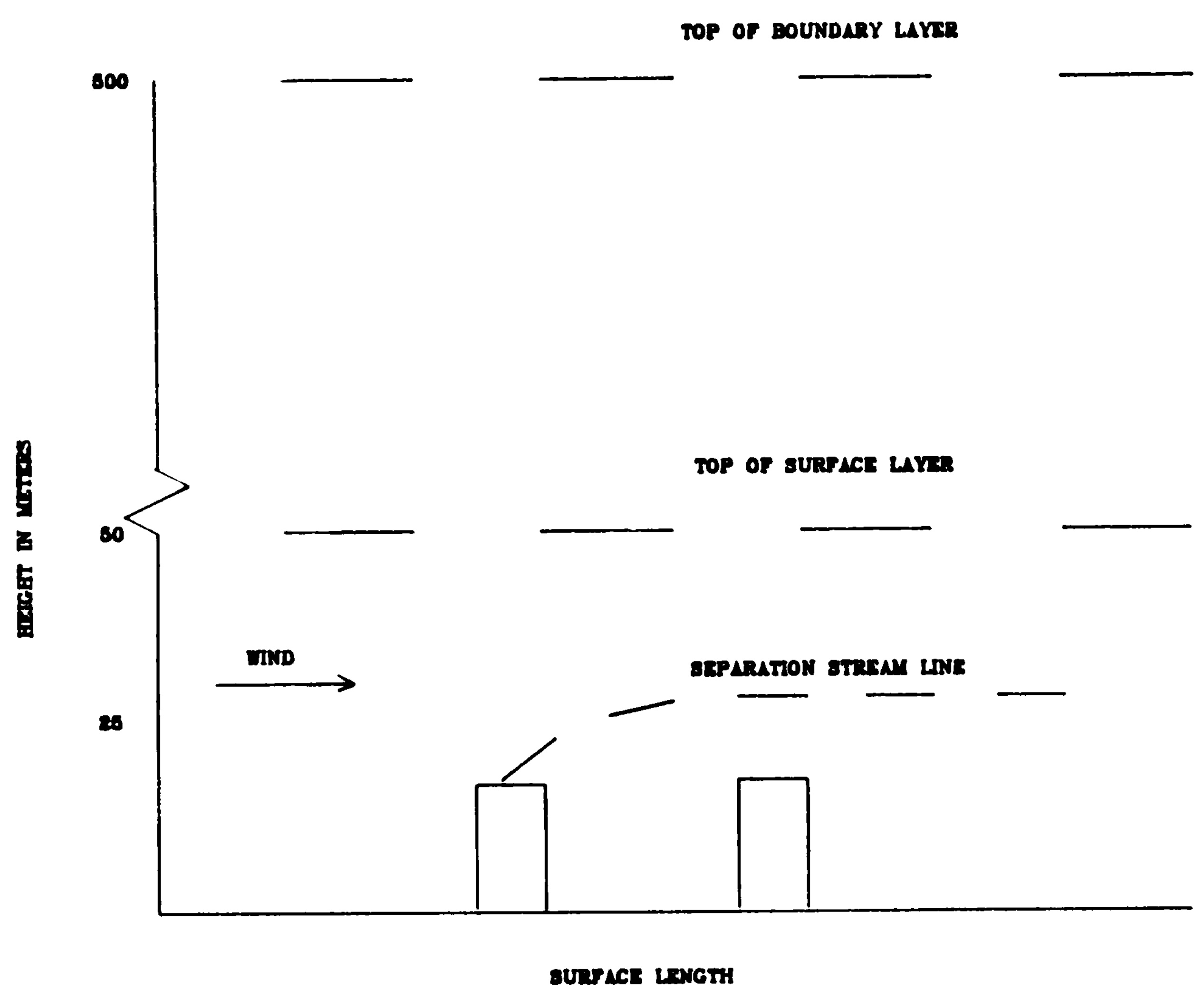


Fig. 1. The surface layer and boundary layer on flat terrain at night time

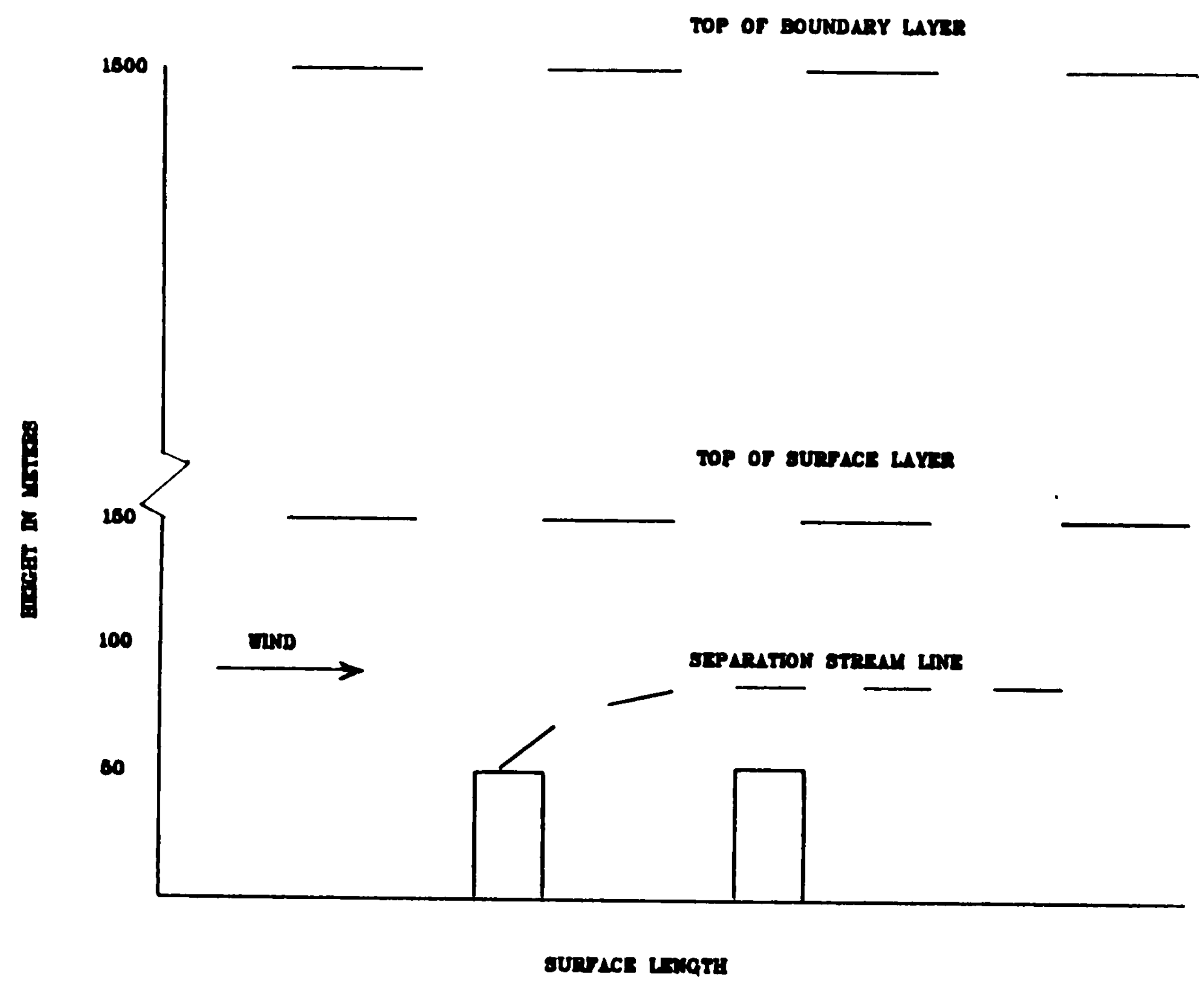


Fig. 2. The surface layer and boundary layer on flat terrain in the day time

boundary layer. Near the land surface, the vertical fluxes remain nearly constant with height (Kaplan and Dinar, 1984).

In addition to mechanical turbulence in this layer, air temperature gradients can alter the wind profile shape. Where temperature gradient does not cause convection, a neutral condition exists and the turbulence is completely mechanical (Panofsky and Dutton, 1984).

When the surface is cool and the air settles to the surface, the condition is defined as stable. In contrast to the stable condition, an unstable condition can occur when the surface is warm and air rises.

These conditions change from day to night. As the sun sets, the temperature of the ground falls rapidly below that of the air, with the result that the layers of the atmosphere immediately in contact with the ground are chilled and become denser than those above. Turbulent flow implies that a mass of air is continually being moved in the vertical direction. When there is a fall in density with a change in height, energy is lost in lifting the denser masses against the force of gravitation, hence, there is less turbulence (Sutton, 1953).

A common configuration of the day and night time boundary layer over rough terrain is shown in Figure 3.

This study will focus on characterizing the wind profile and the surface layer for neutral conditions.

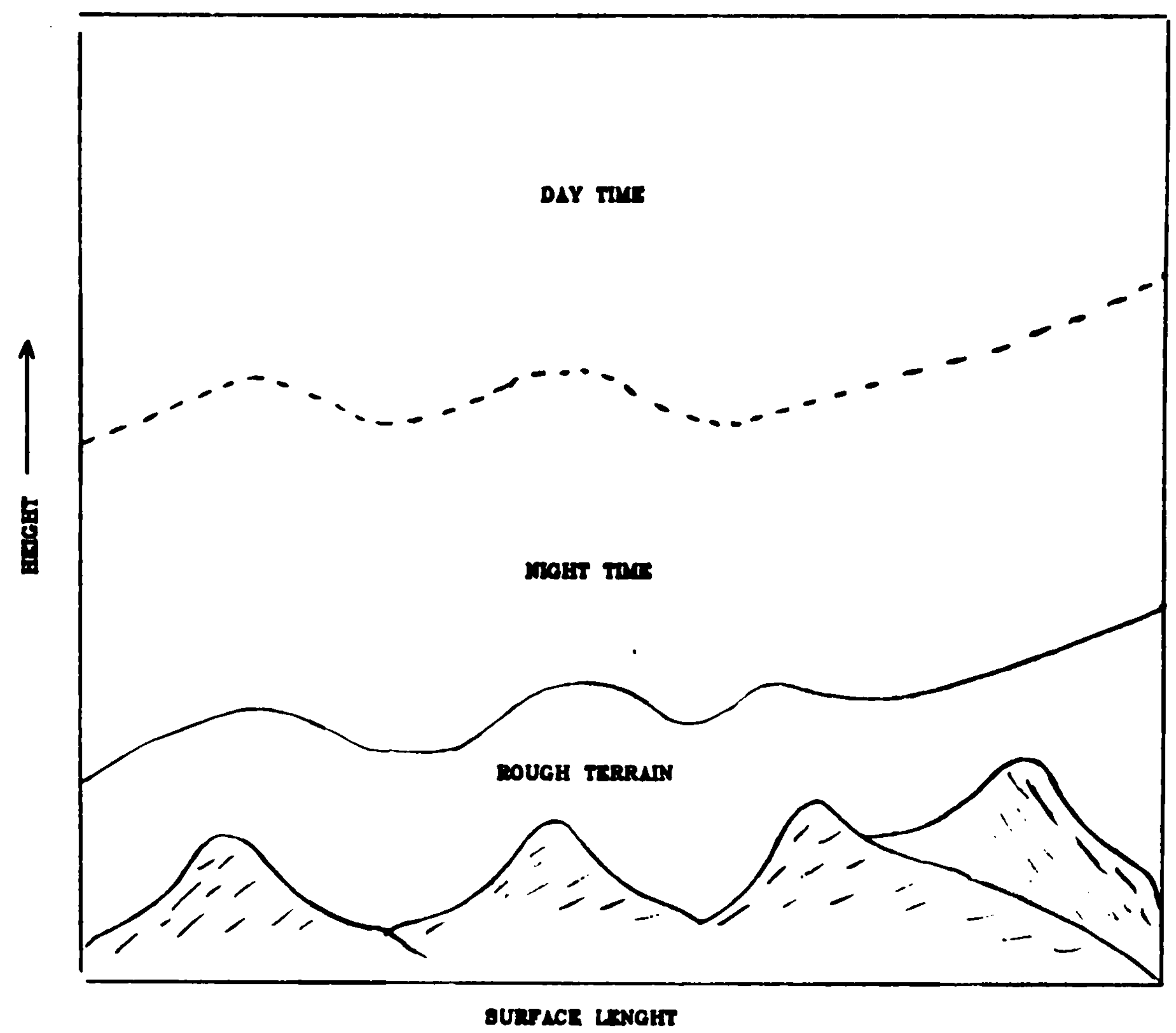


Fig. 3. Typical configuration of the boundary layer for day time and night time over irregular terrain

CHAPTER II

LITERATURE REVIEW

Generally, the wind profile in the atmospheric boundary layer does not follow similarity rules and is often very irregular. Sometimes a maximum wind velocity is measured at the lower half of the boundary layer. Often, observations show irregular change of wind direction with height with no resemblance to the Ekman spiral. The Ekman spiral is the condition where the velocity components perpendicular to the isobars and parallel to the isobars change with height as a spiral form (Sutton, 1953). In conditions in which the pressure field is steady or only slowly changing, the mean motion of the air above a certain height conforms to the theoretical geostrophic wind which flows parallel to the isobars. The wind near the surface of the earth is regarded as the geostrophic wind modified by friction and has a component perpendicular to the isobars and which vanishes at great heights (Sutton, 1953).

The profile gets complex due to many possible factors. Some of the factors are: density stratification of the atmosphere, horizontal inhomogeneity of the temperature field creating a thermal wind, the roughness of the underlying surface, irregular density and curvature of the isobars, orthographic effects, and dynamic effects (Monin, 1973).

To reduce the complications, a simple case of stationary and homogeneous atmospheric boundary layer with rectilinear and evenly distributed isobars is often assumed.

The pressure field can be equated as (Monin, 1973):

$$P(x,y,z) = P_{(z)} + \rho\, f\; G\,(x \sin \alpha - y \cos \alpha) \qquad (2.1)$$

where P = pressure field,

x,y,z = cartesian coordinates; the z-axis is along the vertical and the x-axis coincides with the direction of the surface wind,

α = the angle between the isobars and the surface wind,

ρ = the standard average air density in the layer,

G = the geostrophic wind velocity (this is the velocity of the undisturbed wind on the top of the boundary layer), and

f = frequency.

The Reynold's equations for the mean motion are as follows (Monin, 1973):

$$f(\bar{V} - G \sin \alpha) + \frac{d}{dz}\left[\frac{\tau xz}{\rho} + \nu \frac{d\bar{U}}{dz}\right] = 0 \qquad (2.2)$$

$$-f(\bar{U} - G \cos \alpha) + \frac{d}{dz}\left[\frac{\tau yz}{\rho} + \nu \frac{d\bar{V}}{dz}\right] = 0 \qquad (2.3)$$

where $\overline{U}, \overline{V}$ are the horizontal and vertical components of the mean wind velocity and ν is the coefficient of molecular viscosity of air (Kinematic viscosity).

In the simple case of a horizontal flow whose mean velocity, $\overline{U}$, increases upwards, at the same time there are vertical and horizontal fluctions (W' and U'). The amount of momentum, additional to that of the mean motion at the level in question, carried upwards by an eddy whose fluctuations of velocity are U' and W', is $\rho U'$ per unit volume. There is an assumption of no fluctuation in air density. The rate at which the momentum is carried upwards with velocity, W', is U'W' per unit volume. The average value, $\rho\overline{U'W'}$, is the stress experienced by the mean flow (Scorer, 1958). The same holds true for the YZ plane with velocity fluctuations W' and V'.

$$\tau xz = - \rho \overline{U' W'} \tag{2.4}$$

$$\tau yz = - \rho \overline{V' W'} \tag{2.5}$$

where τxz, τyz = components of the vertical turbulent momentum flux (Reynold's Stresses),

W' = fluctuation of the vertical velocity,

"-" = averaging, and

"'" = there is fluctuation.

Because the x-axis coincides with the direction of the surface wind, the following conditions exist:

$$\frac{\tau yz}{\rho} + \nu \frac{d\bar{V}}{dz} \text{ vanishes at the surface, and}$$

$$\frac{\tau xz}{\rho} + \nu \frac{d\bar{U}}{dz} \text{ approaches some positive limit } U_*^2.$$

where U_* is called the friction velocity.

As shown on Figure 1 and 2, the lower part of the atmospheric layer is the surface layer whose thickness is in tens of meters. Because it is relatively close to the surface of the earth, the action of the coriolis force can be neglected. This surface layer resembles the wall region of boundary layers in nonrotating stratified fluid.

In this part of the atmosphere (the surface layer), the vertical heat and humidity fluxes are small and for neutral conditions we can consider the vertical momentum flux components only (Pasquill, 1974). This results in the following simplification of the governing equations:

$$\frac{\tau xz}{\rho} + \nu \frac{d\bar{U}}{dz} \approx U_*^2 \qquad (2.6)$$

$$\frac{\tau yz}{\rho} + \nu \frac{d\bar{V}}{dz} \approx 0 \tag{2.7}$$

If the surface is rough, the layer near that surface is a dynamic sublayer whose properties are determined by ν, U_* and the roughness (h_s) of the underlaying surface.

If $h_s \leq \nu/U_*$, the roughness does not affect the dynamic sublayer and the surface is called dynamically smooth. Very near a smooth surface a viscous sublayer is formed where the Reynold Stresses are small compared with viscous stresses. The following simplification results for equation 2.6:

$$\frac{\tau xz}{\rho} \approx 0 \tag{2.8}$$

$$\nu \frac{d\bar{U}}{dz} \approx U_*^2 \tag{2.9}$$

Integration of equation 2.9 gives the following function for $\bar{U}$:

$$\bar{U} = \frac{U_*^2 \, z}{\nu} \tag{2.10}$$

This is a linear wind velocity profile over smooth surface. It holds true up to a very small height from the surface.

When $h_s \gg \nu/U_*$, the flow near the surface consists of vortices formed around roughness elements and no viscous sublayer exists. The surface is called dynamically rough. The land surface is always dynamically rough. Therefore, the wind velocity profile over land surfaces should be different from the above linear profile. In this case, the viscous stress is small compared to the Reynold Stress (Monin, 1973). For this condition, the following simplification results:

$$\nu \frac{d\bar{U}}{dz} \approx 0 \tag{2.11}$$

$$\frac{\tau xz}{\rho} = U_*^2 \tag{2.12}$$

τxz will be referred as τ from now on and rearrangement gives the following function for the friction velocity:

$$U_* = \sqrt{\frac{\tau}{\rho}} \tag{2.13}$$

Because of the chaotic nature of turbulent flow, shear can not be determined in a simple analytical way as is the case of laminar flow. Researchers have used statistical methods to develop the relationship between shear on the boundary and the velocity of fluid. This shear is a measure of the momentum transfer between adjacent layers of air. Prandtl has suggested the following equation for shear (Albertson, et al., 1960):

$$\tau = l^2 \left(\frac{du}{dz}\right)^2 \tag{2.14}$$

where l is the scale of turbulence and is determined experimentally.

Prandtl further assumed that l is proportional to z and developed the following equation:

$$\tau = \rho\, k^2\, z^2 \left(\frac{du}{dz}\right)^2 \tag{2.15}$$

where k is Von Karman's Constant and is equal to 0.4 for neutral atmospheric conditions.

$$\frac{\tau}{\rho} = k^2\, z^2 \left(\frac{du}{dz}\right)^2 \tag{2.16}$$

Taking the square root of both sides, equation 2.16 becomes:

$$\sqrt{\frac{\tau}{\rho}} = k\, z\, \frac{du}{dz} \tag{2.17}$$

Because $\sqrt{\frac{\tau}{\rho}} = U_*$,

$$U_* = k\ z\ \frac{du}{dz} \qquad (2.18)$$

$$du = \frac{U_*}{k}\ \frac{dz}{z} \qquad (2.19)$$

Integration from the lower limits of zero for velocity and Z_o for height to the upper limits of U for velocity and Z for height gives:

$$\int_0^U du = \frac{U_*}{k} \int_{Z_o}^{Z} \frac{dz}{z} \qquad (2.20)$$

$$U = \frac{U_*}{k} \ln \frac{Z}{Z_o} \qquad (2.21)$$

It should be noted that Z_o is the height at which U is zero. It is a boundary condition and should be a function of the surface or boundary over which the wind moves. This value is defined as the aerodynamic roughness.

When wind blows over surfaces with roughness objects, the profile is displaced or raised to a new reference plane. To obtain a reasonable agreement between actual wind profiles and the logarithmic profile described by equation 2.21, it is necessary to raise the profile from the ground surface by a height D which is defined as the displacement height as shown on Figure 4.

The displacement height D implies a shift in height scales by an amount D (Lowry, 1967). In another definition, the displacement height is taken as the depth of the still air trapped among the roughness elements of the surface (Sutton, 1949). With this modification, the wind profile equation becomes:

$$U = \frac{U_*}{k} \ln \left(\frac{Z - D}{Z_o}\right) \tag{2.22}$$

The estimation of the displacement height D and the aerodynamic roughness Z_o has been a perplexing problem. Various researchers (Schwab et al., 1981 and Lyles and Allison, 1979) have defined these parameters differently and offered different equations to estimate them. There appears to be no universal prediction equations for the displacement height or aerodynamic roughness.

Some of the estimation equations of the aerodynamic roughness Z_o from various literature sources are given in Table 1.

The aerodynamic roughness parameter, Z_o, is physically represented by the distance from the displaced reference plane to the height where the wind profile extrapolates to zero. It has been suggested that Z_o represents the eddy size at the surface (Panofsky and Dulton, 1984). The rougher the surface, the larger the eddies.

From this literature review it was found that displacement height lacks a precise definition. It was also found that the aerodynamic

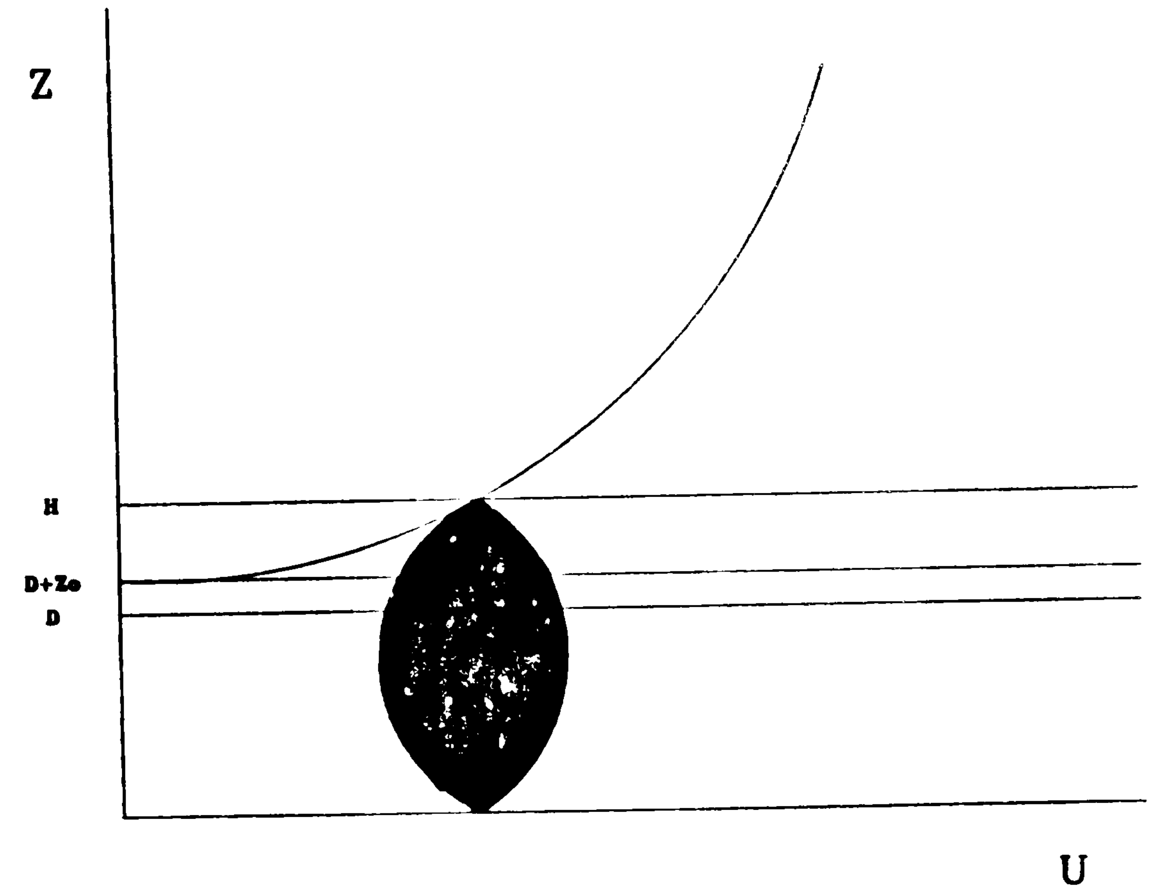


Fig. 4. Typical wind profile over an object

Table 1. The estimation equation of various researchers for the aerodynamic roughness parameter, Z_o

EQUATIONS	SOURCE
$Z_o = h - d - E$	Lyles and Allison, 1979
$\log Z_o = a + b \log H$	Seginer, 1974
$Z_o = 0.04H^{1.417}$	Sellers, 1965
$Z_o = 0.13H^{0.997}$	Tanner and Pelton, 1960
$Z_o = 0.06H^{1.19}$	Kung, 1961, 1963
$Z_o = 0.1H$ (grass)	Saxton, et al., 1974
$Z_o = 0.025H^{1.1}$ (corn)	Saxton, et al., 1974
$Z_o = (1.08 \frac{A_r}{A} - 0.08)H^1$ for $0.11 \le \frac{A_r}{A} \le 0.25$	Counhan, 1971
$Z_o = 0.5H^* s/S$	Lettau, 1969
$Z_o = 0.033\ d$	Nikuradse, 1950
$Z_o = 0.1H$ and $D = 0.64H$	Houghton, 1985
$Z_o = 0.056\ H_c 1.37$	Houghton, 1985
$Z_o = -D$	Sutton, 1949
$Z_o = 0.1\ (Z - D)$	Monteith, 1975
$Z_o = \frac{10}{U}$	Ishizaki, 1983

h = roughness height,
d = "effective" height of roughness,
E = error in origin. The distance below the roughness height (h) to the point where U = 0,
H = vegetative height in cm,
A_r = the plan surface area of roughness elements,
A = the total plan surface area,
H^1 = average roughness height,
H^* = effective obstacle height in cm,
s = the silhouette area (cm^2) of the average obstacle or area measured in the vertical crosswind-lateral plane,
S = A/N (specific area),
A = total area in the horizontal plane,
N = total number of roughness elements,
d = particle diameter
H_c = average height of roughness in meters, and
D = displacement height.

roughness term has more than one definition. The most common definition used for Z_o is the height where the wind profile extrapolates to zero after the profile has been adjusted for the displacement height. The literature review also revealed numerous empirical equations for estimating Z_o. It is unknown which equation, if any, is universal in nature.

Based on the literature review, the objectives listed in Chapter I need to be researched.

CHAPTER III

DEVELOPMENT OF PREDICTION EQUATIONS

The review of literature for the estimation of displacement height, D, and aerodynamic roughness, Z_o, revealed that ambiguity exists in the methods used to estimate these two variables. Mathematical and physical explanation of these parameters is not complete.

The development for displacement height will be presented first.

Displacement Height D

As defined earlier, displacement height is the depth of still air trapped among the roughness elements (Sutton, 1949). It is also the amount of shift in height scale necessary to make the logarithmic profile equation work (Lowry, 1967). Both definitions can be illustrated by visualizing a smooth, rigid plastic cloth a distance D above the soil surface. Above the cloth, the logarithmic profile will develop, but below the cloth, still air will exist. The distance from the soil to the cloth is the average height of the surface.
Using this definition, the following equation can be written for displacement height:

$$D = H_c \; F_c \qquad (3.1)$$

where H_c = average height of an individual roughness element as viewed from above, and

F_c = fraction of the total surface covered by roughness elements.

Obviously H_c is a function of element shape and maximum height. In addition, the rigidity of the roughness object will affect H_c since high winds can cause non-rigid elements to bend over and have less height. While it is difficult to estimate H_c for non-rigid crops that wave in the wind, it is possible to estimate both H_c and F_c for several rigid cover conditions.

A single layer of spheres representing sand or clods is relatively easy to evaluate. First, these objects can assume two possible configurations (open pack or closed pack) to cover the surface. These packing arrangements represent the lower and upper boundary conditions for fraction of surface covered. The actual cover for a random arrangement of spheres should lie between these boundary conditions.

The open pack arrangement is shown in Figure 5. The top view shows the possible contribution of each sphere to cover the area. The ratio of the area covered to the area of the square ABCD is the fraction of cover, F_c. Also see Figure 5. The total area is just the area of the square (d^2, where d is the diameter of the sphere). The covered area is the sum of the four quarters of the circle ($\pi\ d^2/4$). The ratio of these two areas has the value of $\pi/4$ or 0.785.

As shown in Figure 5, the average height (H_c), as seen from above, is the average elevation of the top one-half of the sphere, which can be predicted with the following equation:

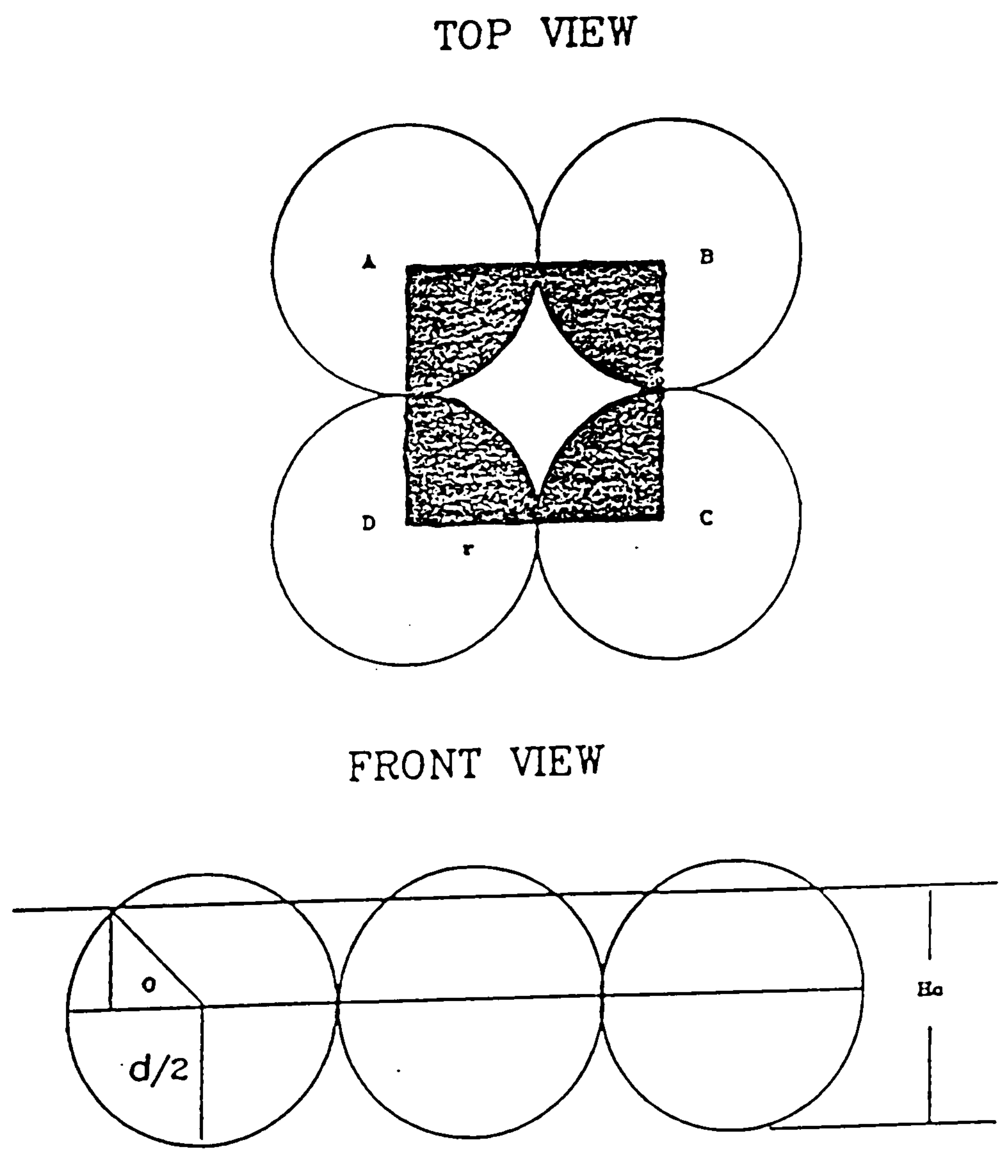


Fig. 5. Open pack arrangement

$$H_c = \frac{d}{2} + \frac{d}{2} \sin 45^o \tag{3.2}$$

which reduces to:

$$H_c = 0.85\ d \tag{3.3}$$

Using the definition of displacement height, the value for D for spheres in an open pack arrangement is:

$$D = 0.85\ d \times 0.785 \tag{3.4}$$

which reduces to:

$$D = 0.67\ d \tag{3.5}$$

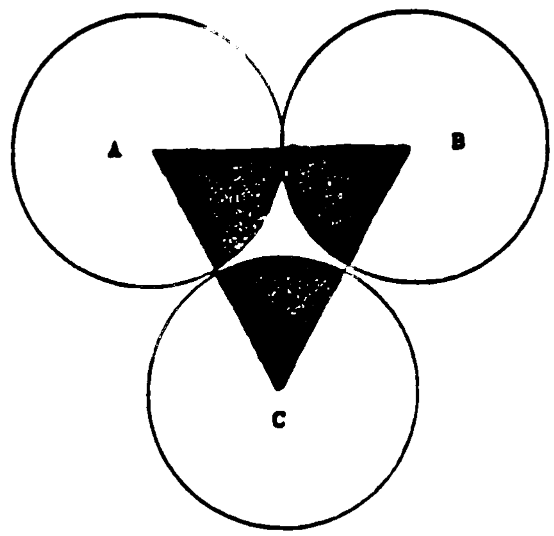
This represents the lower boundary condition for displacement height.

The second possible configuration is the closed pack arrangement. As shown in Figure 6, the ratio of the shaded area to the area of the triangle ABC gives the fraction of cover. The total area is given as:

$$\text{Total area} = \frac{1}{2} d^2 \sin 60 \tag{3.6a}$$

$$= 0.433\ d^2$$

TOP VIEW



FRONT VIEW

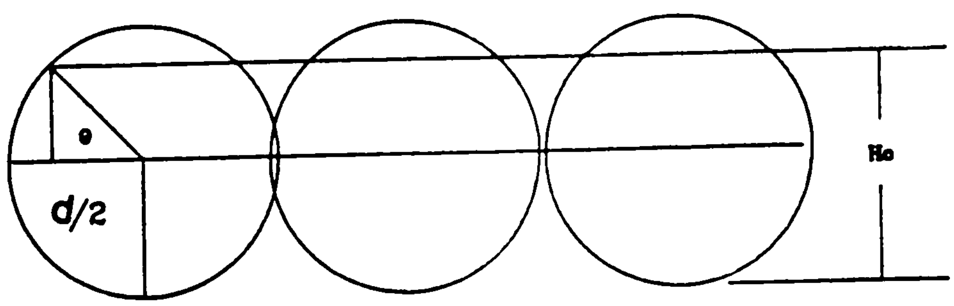


Fig. 6. Closed pack arrangement

The equation for the covered area is:

$$\text{Covered area} = 3\left(\frac{1}{6}\,\pi\,\frac{d^2}{4}\right) \qquad (3.6b)$$

$$= 0.393\ d^2$$

The value for the ratio of covered area to total area is 0.91.

The average element height H_c is 0.85 d as shown earlier. The displacement height for the closed pack configuration is given as follows:

$$D = 0.85\ d \times 0.91 \qquad (3.7a)$$

which reduces to:

$$D = 0.77\ d \qquad (3.7b)$$

This is the upper boundary condition for displacement height.

Averaging the displacement height for the two configurations results in an equation to estimate the displacement height for randomly arranged spheres and objects such as sand and clods,

$$D = \frac{1}{2}\,(0.67 + 0.77)d \qquad (3.8)$$

$$D = 0.72\ d \qquad (3.9)$$

It is also relatively easy to estimate displacement height for ridges. Consider, for example, triangular shaped ridges with bases that touch. The average height of the element as seen from above is just 0.5 times the ridge height. The fraction of cover is 1.0 so the displacement height is given by the following equation:

$$D = 0.5 \ H \tag{3.10}$$

where H = ridge height.

If the wind is parallel with the ridges, then the displacement height is computed using the clod roughness but not the ridge height.

The estimation of displacement height for vegetation can be more complex. For fruit trees such as apples, oranges, and peaches, the open pack arrangement for spheres should be a good estimate. This compares closely with the regression equation of Stanhill (1969) for vegetation in general (see Figure 7).

$$D = 0.7 \ H \tag{3.11}$$

At the one digit accuracy level, both equations are the same.

For immature orchards, where the trees are too small to touch, the fraction of cover will have to be measured. For this condition, the following equation should be used:

$$D = 0.85 \ H \ F_c \tag{3.12}$$

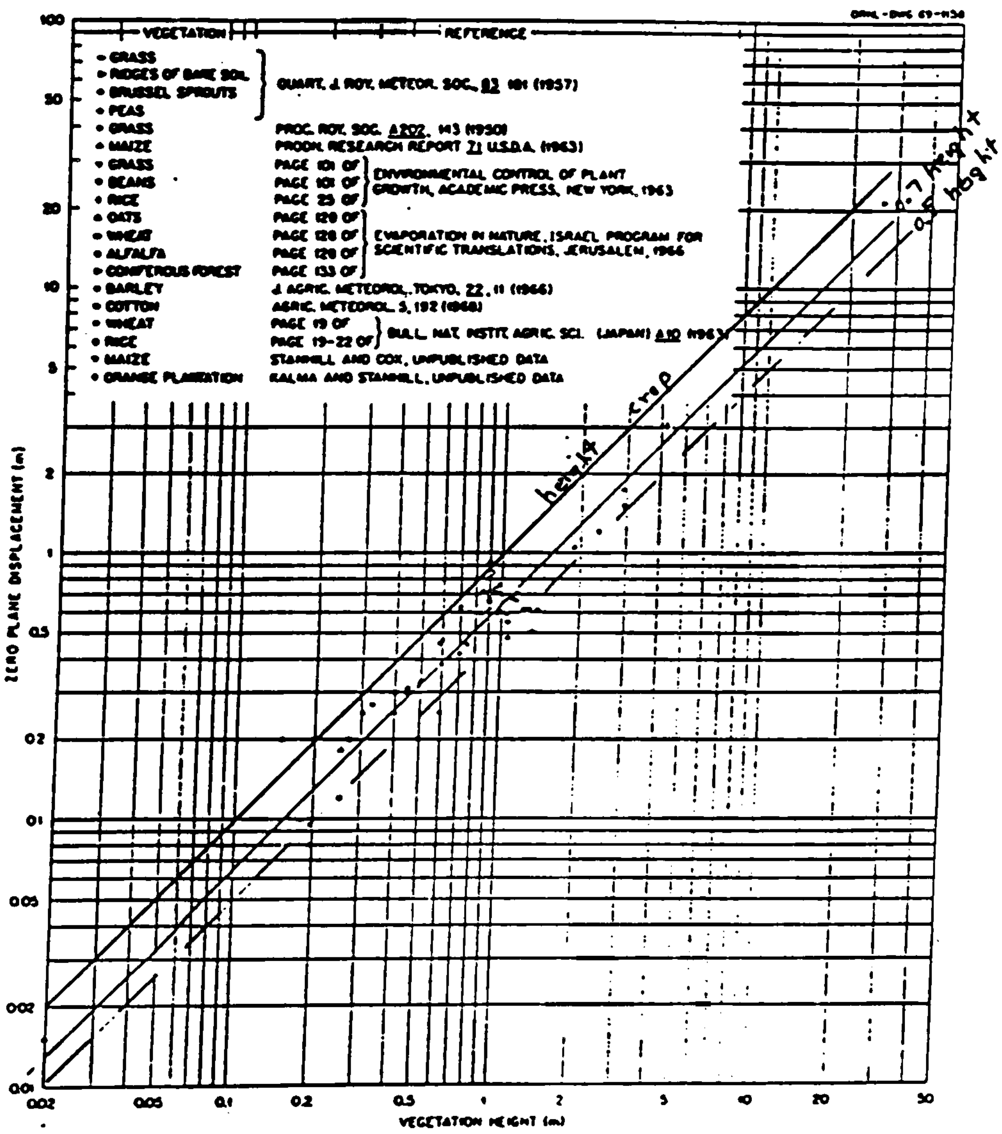


Fig. 7. Relationship between zero-plane displacement D and vegetation height H for different vegetation types (Stanhill, 1969)

This equation should also be a good estimator for displacement height of most row crops such as soybeans, cotton, sunflower, or corn. Most field crops in well watered climates will exhibit between 0.75 and 0.95 surface of cover. This results in a 0.64 to 0.81 range for the coefficient times the crop height. Again, these are reasonably close to the results of Stanhill (1969) (see Figure 7). The lower value exactly matches the estimation suggested by Houghton (1985),

$$D = 0.64\ H \tag{3.13}$$

The value of using the definition to estimate displacement height instead of the regression equation reported by Stanhill (1969) is that individual plant shape and arrangement can be considered.

Aerodynamic Roughness, Z_o

The roughness parameter (Z_o) and the zero-plane displacement (D) are structurally based parameters and are assumed not to depend on wind velocity (Houghton, 1985). As shown in the previous chapter, various attempts have been made to express these parameters in different empirical equations.

The aerodynamic roughness parameter (Z_o) is dependent on the type of surface and may depend on atmospheric stability. Sutton (1949) has suggested that it is a function of the height of roughness, the displacement height, and the atmospheric conditions,

$$Z_o = f(H,\ D,\ A_s) \tag{3.14}$$

where A_s is the atmospheric stability factor.

The following Z_o and K values developed over short grassed surfaces have been suggested (Sutton, 1949) for variations in temperature gradient.

Temperature Gradient	Z_o(cm)	K
Super adiabatic lapse	0.006	0.61
Adiabatic gradient	0.28	0.40
Inversion	0.80	0.22

An alternative view of the effect of atmospheric stability on the wind profile is given as follows (Hagen, 1980):

$$U = \frac{U_*}{K} \ln \left(\frac{Z - D}{Z_o}\right) + f(Ri) \tag{3.15}$$

where U is mean velocity at height Z and f(Ri) indicates that mean velocity also depends on Richardson's number (Ri), which is a measure of atmospheric stability. Under neutral stability conditions, f(Ri) is zero. By specifying the velocity at a height where Z = H, the number of unknowns can be decreased by one,

$$U_H = \frac{U_*}{K} \ln \left(\frac{H - D}{Z_o}\right) \tag{3.16a}$$

For neutral conditions, this equation reduces to:

$$U_H = \frac{U_*}{0.4} \ln \left(\frac{H - D}{Z_o}\right) \tag{3.16b}$$

and the present work will be restricted to neutral conditions.

For the neutral atmospheric condition, the logarithmic equation 3.16b can be reformulated to make Z_o the subject of the equation,

$$Z_o = [e^{-0.4 \frac{U_H}{U_*}}] (H - D) \tag{3.17}$$

Pasquill (1974) has given typical values of $U_*/U_{(10)}$ for various surfaces. When the roughness height is 10 meters (forests and urban areas), $U_*/U_{(10)}$ becomes an estimate for U_*/U_H. He has given a constant value of 0.17 for neutral conditions. From this reference, it can be concluded that U_H/U_* is a constant value for a given atmospheric stability condition.

This supports the view that Z_o is not dependent on U and U_*. Therefore:

$$Z_o = A(H - D) \tag{3.18}$$

where A is a constant.

While the value for U_*/U_H from Pasquill (1974) is only one point and a rough estimate, it does provide an estimate for A. The A value that is obtained is 0.1.

From literature sources (Armbrust et al., 1964), nineteen wind profiles over spherical objects, ridges, and sorghum residue were obtained to get measured values of wind velocity at (U_H), and shear velocity U_*. The displacement height (D) was estimated using equation 3.9.

The aerodynamic roughness parameter Z_o, was calculated using U_H, U_*, H, and D. In some cases, the data was given for the complete profile and Z_o could be determined directly by fitting equation 2.23 to the data.

Linear regression was used with this data to obtain the following equation for Z_o:

$$Z_o = 0.13\ (H - D) \qquad (3.19)$$

This equation had an R^2 value of 0.92 and was significant at the 99.9 percent level ($\alpha = .001$). This relation is shown in Figure 8.

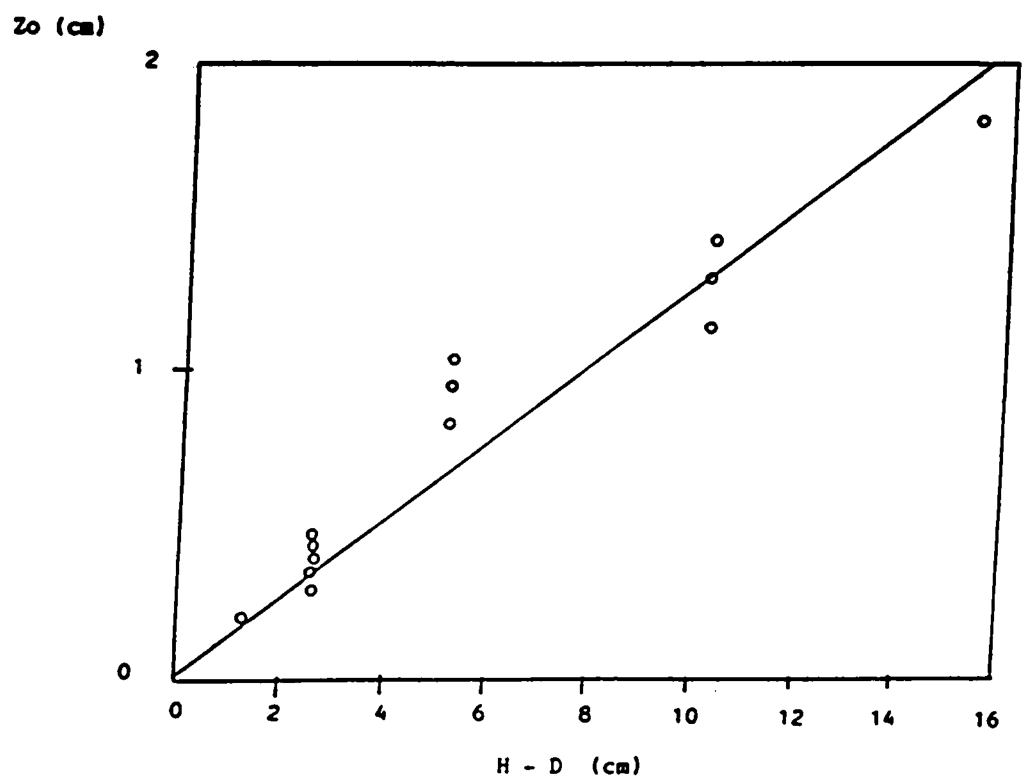


Fig. 8. Aerodynamic roughness height Z_o as a function of (H - D) for nineteen wind profile measurements (Data analyzed from Armbrust et al., 1964)

CHAPTER IV

DISCUSSION AND FURTHER VERIFICATION OF DEVELOPED EQUATIONS

It is important to note that accurate prediction of Z_o with equation 3.19 requires an accurate prediction of displacement height. This would explain why numerous equations have been developed for various sized and shaped roughness elements. Without displacement height in the equation, it is impossible to predict Z_o over a wind range of roughness height. The derived equation for Z_o is unique in that it includes both height of the elements and the displacement height. Because it includes displacement height, it will be shown that this equation is a universal equation which accurately predicts Z_o over all heights of interest.

First, it will be shown that the equation in Table 1 obtained by Nikuradse (1950) for sand grain roughness is a subset of equation 3.19. When equation 3.19 is restricted to a surface of spheres, H becomes the diameter of the sphere and D can be estimated with equation 3.9. For this restruction, equation 3.19 reduces to:

$$Z_o = 0.036\ d \qquad (4.1)$$

Table 2 shows a comparison of Z_o values calculated by equation (3.19), Nikuradse's equation, and values derived from velocity profiles by various researchers.

Table 2. Comparison of Z_o values calculated by the equation $Z_o = 0.13\ (H - \bar{D})$; Nikuradse's equation and values derived from mean velocity profiles by various researchers

KIND OF PARTICLE	d, CM	Z_o (NIKURADSE), CM	$Z_o = .13*.28d$, CM	Z_o PROFILE, CM	SOURCE
		$Z_o = .033d$			
Spheres	0.41	0.0137	0.0149	0.0080	Schlichting (1960)
Sph. (Tapeoca)	0.61	0.0202	0.0222	0.0189	Lyles et al. (1971)
Sph. (Glass)	1.64	0.0547	0.0597	0.0499	Lyles et al. (1971)
Sph. (Glass)	2.45	0.0818	0.0892	0.0988	Lyles et al. (1971)
Gravel	0.238-0.283	0.0087	0.0095	0.0328	Chowdhury (1966)
Gravel	0.200-0.332	0.0089	0.00976	0.0293	Lyles & Allison (1979)
Gravel	0.635-0.952	0.0265	0.0289	0.0543	Lyles & Allison (1979)
Sand (Smoothed)	0.015-0.029	0.0007	0.0008	0.0007-0.00047	Lyles & Allison (1979)
Sand (Smoothed)	0.042-0.059	0.0017	0.0018	0.0009-0.00040	Lyles & Allison (1979)
Sand	0.015-0.025	0.0007	0.0007	0.00046	Zingg (1953)
Sand	0.025-0.030	0.0009	0.0010	0.00137	Zingg (1953)
Sand	0.030-0.042	0.0012	0.0013	0.00274	Zingg (1953)
Sand	0.042-0.059	0.0017	0.0018	0.00366	Zingg (1953)
Sand	0.059-0.084	0.0024	0.0026	0.00487	Zingg (1953)

Nikuradse's equation can also be used to obtain an alternative estimate for A in equation 3.18.

$$Z_o = 0.033d \quad \text{(Nikuradse's equation)}$$
$$Z_o = A \times 0.28d \text{ (equation 3.18 for spheres)}$$
$$0.033d = A \times 0.28d$$
$$A = 0.12 \tag{4.2}$$

At this point, three estimates for A have been obtained: 0.10 from a single point recommendation by Pasquill (1974), 0.12 from the results of Nikuradse (1950), and 0.13 from the data shown in Figure 8. All of these values are relatively close, which suggests that A is truly a constant. These results also suggest that equation 3.19 is a universal equation for Z_o which is an accurate predictor of Z_o for a wide range of heights and surface conditions.

A final comparison of Z_o values calculated with equation (3.19) and ranges of Z_o values reported by various literature sources is given in Table 3 and graphically expressed in Figure 9. Note that equation 3.19 predicts values within the reported ranges from sand grains to cities.

This represents at least four orders of magnitude of roughness height for which the equation appears to be valid. This further supports the universality of equation 3.19.

Because equation 3.19 appears to be a more general equation than Nikuradse's equation, it may provide a useful estimation tool for estimating drag in pipes with surface roughness other than sand grains such as corrugations.

Table 3. Ranges of Z_o values from various literature sources compared to values calculated by the equation $Z_o = 0.13\ (H - D)$

TYPE OF SURFACE	SOURCE	Z_o(cm)	CALCULATED Z_o(cm)	H(cm)
Bare Soil	Lettau, 1969	0.058	0.036	1
Sand	Lettau, 1969	0.0036	0.0036	0.1
Mown Grass	Monin, 1973 Priestly, 1959	0.2-0.7		1.5-3
	Houghton, 1985	0.1-1	0.1	3
Low Grass	Lowry, 1967 Khayarattee, 1985	1-4	1.1	30
Tall Grass	Monin, 1973	4-9		60-70
	Houghton, 1985	3.9		
	Lowry, 1967	4-10		
	Khayarattee, 1985	4-10		
	Sutton, 1953	3	2.9	60
Wheat	Houghton, 1985	4.5	3.97	65
Low Woods	Houghton, 1985	5-10	5.46	150
High Woods	Houghton, 1985	20-90	29	800
Trees	Houghton, 1985	10-30	11	300
Suburbia	Houghton, 1985	100-200		
	Khayarattee, 1985	20-40	30	500
City	Houghton, 1985	100-400		
	Khayarattee, 1985	60-80	120	2000

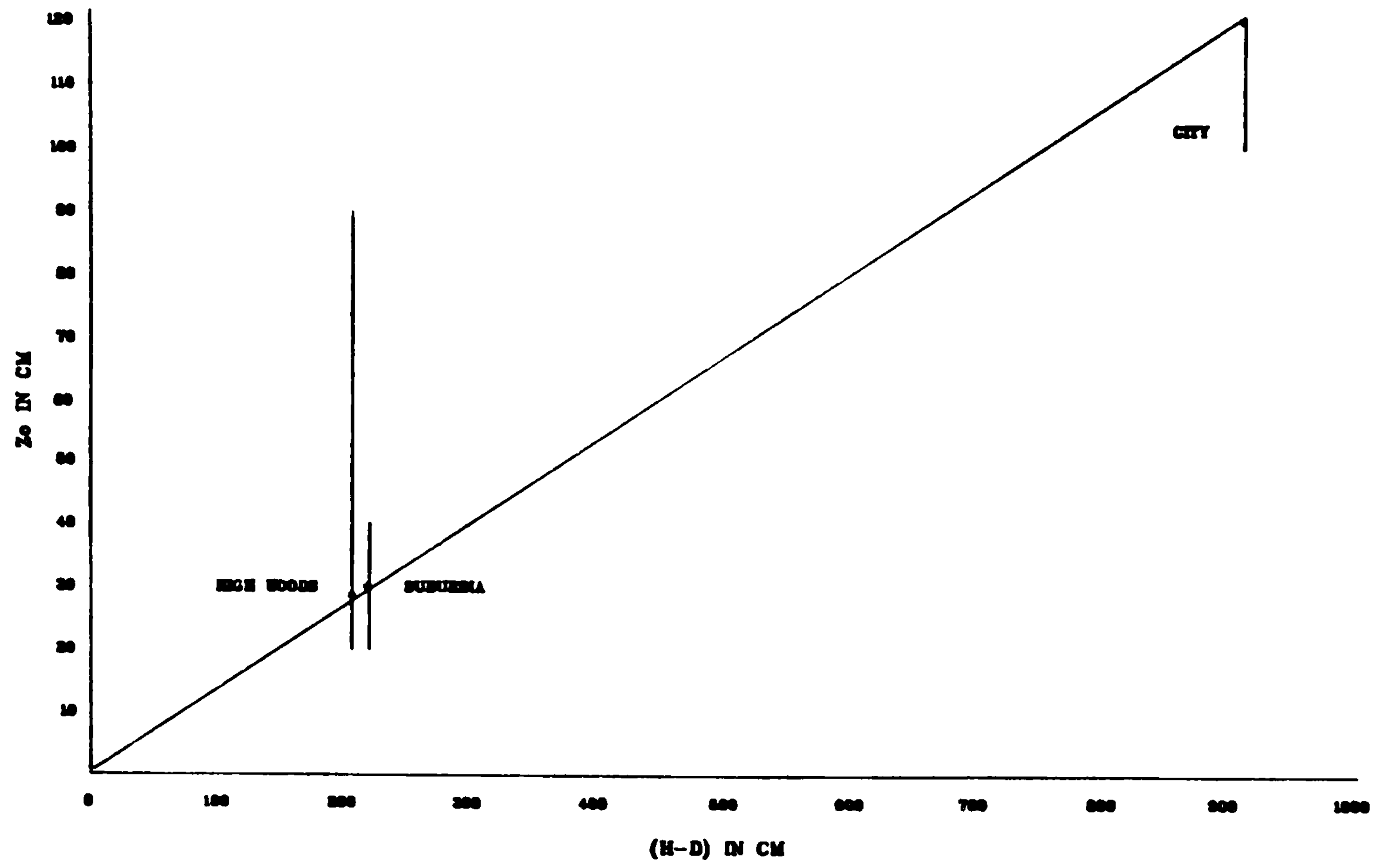


Fig. 9. Comparison of calculated Z_o point values with ranges from literature sources

While equation 3.19 appears to be universal, it does have some limitations. These limitations will now be discussed.

The aerodynamic roughness parameter, Z_o, can only be visualized when the roughness objects are fairly close to each other. If the objects are far apart, Z_o should be that of the surface without the objects.

The lower fraction of cover or maximum spacing for which equation 3.19 is valid will now be evaluated. Greenley and Iversen (1985) have reported values of (1/30)H and (1/8)H for the respective lower and upper boundary conditions for Z_o.

The lower boundary condition is when the objects are close to each other. In case of spheres, it is when D = 0.72 d and H - D = 0.28 d as shown in Figure 10A.

$$Z_o = 0.13\ (0.28\ d) = \frac{1}{27}\ d \qquad (4.3)$$

This closely matches the minimum value given by Greeley and Iversen (1985).

The upper boundary condition is where the maximum Z_o value occurs. A possible configuration is shown in Figure 10B.

This boundary condition can be used to estimate the lower fraction of cover for which equation 3.19 is valid,

$$\frac{H}{8} = .13\ (H - D) \qquad (4.4)$$

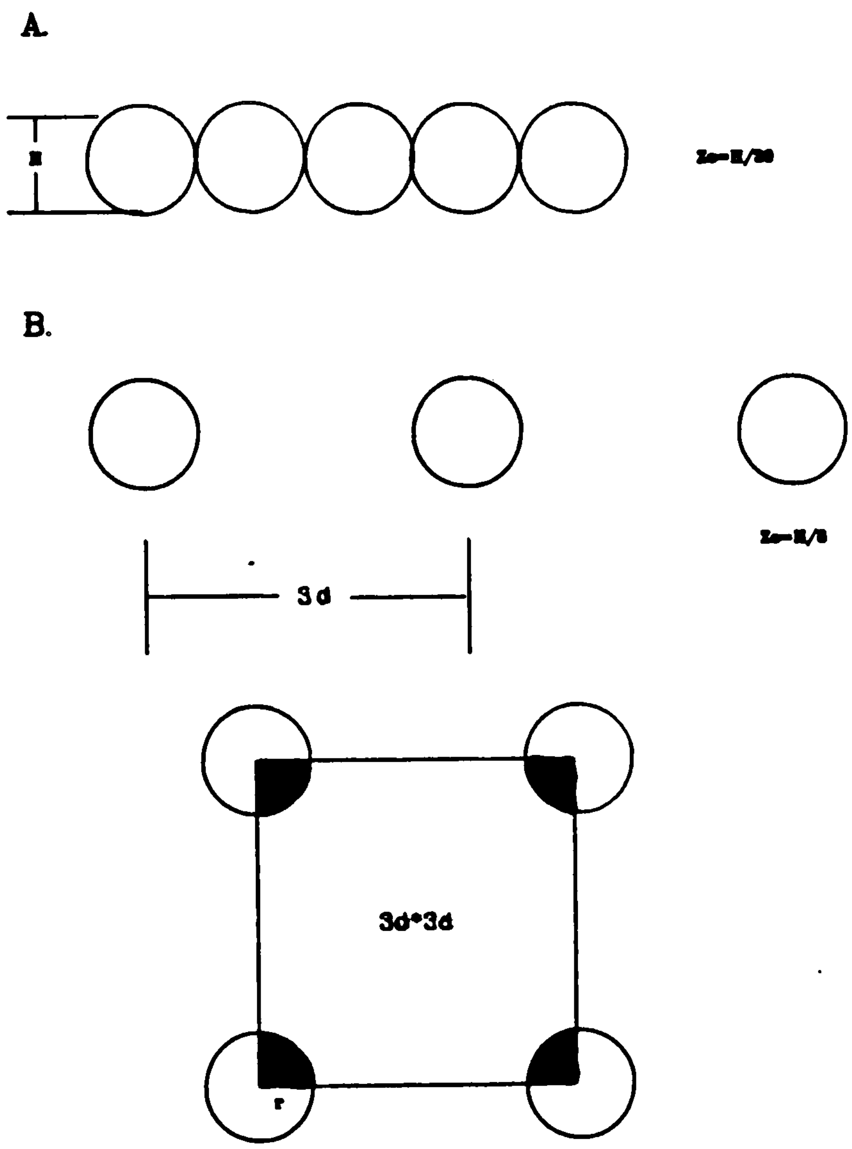


Fig. 10. Configuration of spheres for minimum and maximum Z_o values

The displacement height can be estimated with equation 3.1 where H_c is equal to H,

$$\frac{H}{8} = .13\ (H - H\ F_c) \tag{4.5}$$

Rearrangement and simplification gives the following:

$$0.96 = (1 - F_c) \tag{4.6}$$

which results in a lower limit of 0.04 for the fraction of cover. For an open pack arrangement of spheres, this is equivalent to a spacing of 4.4 diameters of the particle. Iversen (1985) suggested that the maximum Z_o value occurs when the objects are apart by about two diameters. This would give a lower value of 0.20 for fraction of cover. An average value of three particle diameters and a fraction of cover of 0.1 is probably a reasonable lower limit for which equation 3.19 is valid.

Another limitation on equation 3.19 is that it was developed for neutral conditions. As shown in equation (3.18), $e^{-0.4 \frac{U_H}{U_*}}$ is a constant and is estimated to be 0.13 for neutral atmospheric conditions.

$$e^{-0.4 \frac{U_H}{U_*}} = 0.13$$

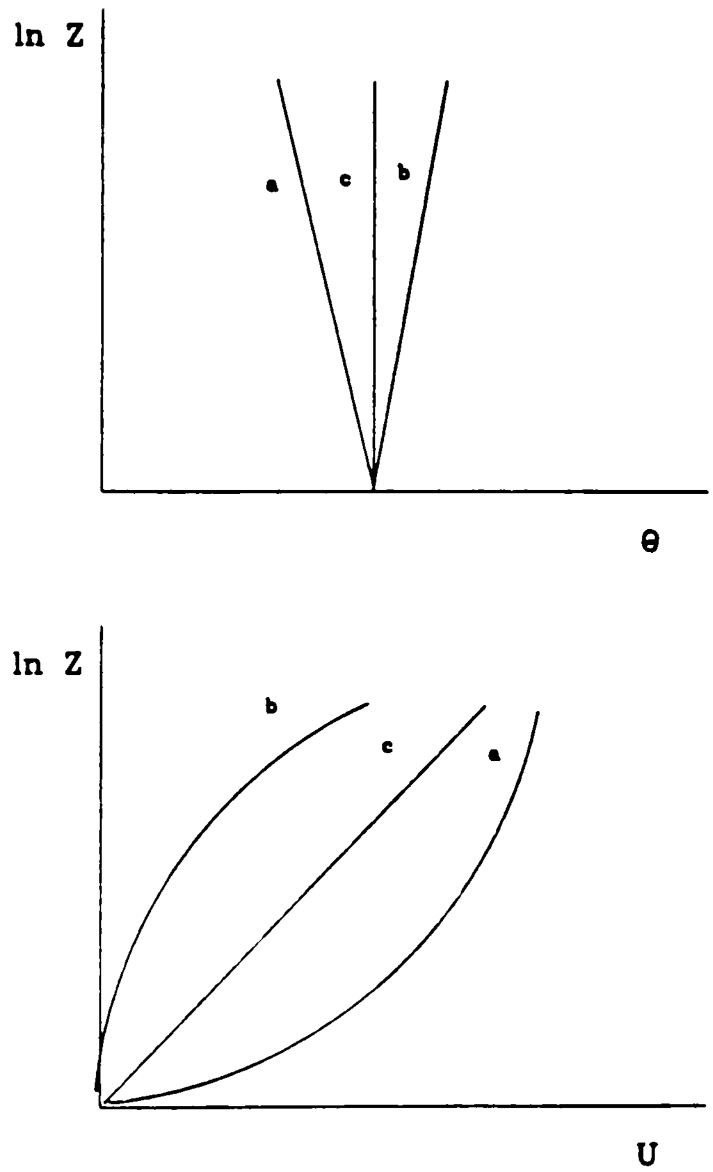
$$\frac{U_H}{U_*} = \ln \frac{\frac{1}{(0.13)}}{0.4}$$

$$= 5.1 \qquad (4.7)$$

It can be concluded that the ratio of the average velocity, $\bar{U}$, measured at a height Z = H is about five times as much as the shear velocity, U_*. Pasquill (1974) has given U_H/U_* a value of 5.88 for a flow over 10 meters high forests and urban areas. Pasquill also reported that the value of U_H/U_* changes as atmospheric stability changes. This will cause a change in the A value of equation 3.18 (see equation 3.17).

The variation of temperature, Θ, and wind speed, U, with height for three different atmospheric conditions is shown in Figure 11 (Mieghem, 1973). The gradient of the log height versus average velocity line increases from neutral to inversion conditions. U_* decreases when the gradient increases. U_H/U_* should be higher than 5.1 for inversion conditions. It can be postulated that the same velocity wind is more erosive under neutral conditions than inversion conditions.

The log height versus velocity gradient is smaller for lapse rate conditions than for neutral conditions. When the gradient is smaller, U_* increases. U_H/U_* should be smaller for lapse rate conditions than for neutral atmospheric conditions. This conclusion matches with



(a) LAPSE RATE CONDITIONS

(b) INVERSION CONDITIONS

(c) NEUTRAL CONDITIONS

Fig. 11. Log height versus temperature and wind speed

Pasquill's (1974) values. Again, it can be concluded that the same speed wind is less erosive under neutral conditions than lapse rate conditions.

CHAPTER V

CONCLUSION

Wind is a climatic variable that has major impact on surface conditions on the earth. Both soil erosion and evapotranspiration are examples of surface phenomena that are highly sensitive to the shape of the wind profile. If these processes are to be accurately predicted, the wind profile parameters must be measured or predicted accurately.

The objectives of this study were: 1) to physically define the displacement height, D, 2) to develop a prediction equation for displacement height, 3) to develop a prediction equation for aerodynamic roughness, and 4) to verify the developed equations using published data.

Displacement height was defined as the average height of the total surface roughness. Based on this definition, an equation was developed to predict displacement height as a product of the average height of each roughness element and the fraction of surface covered by the roughness element. The average height of each object varies with the shape of the object and the fraction of cover varies with the packing arrangement of the objects and the number of objects. Specific evaluations were made for an open pack arrangement of spheres, a closed pack arrangement of spheres, and the average of these two arrangements.

The equation developed for an open pack arrangement of spheres completely agrees with the relationship $d = 0.64\ H$ derived by Cowan

(1968). The equation which resulted from the average of open and closed pack arrangements (D = 0.72 H) closely matched the equation obtained by regression (D = 0.7 H) reported by Stanhill (1969).

It was found that accurate knowledge of displacement height was an important parameter in the prediction of aerodynamic roughness. Most equations for aerodynamic roughness that are found in the literature do not include displacement height and therefore, are not universal. In this study, an equation for Zo was derived by rearrangement of the wind profile equation. In this process, it was noted that the ratio of wind velocity at a specific height to the U_* value is a constant for a given condition of atmospheric stability (neutral, stable, unstable). For neutral atmospheric conditions, it was found that predictions with the derived equation for Z_o matched published measurements and estimates over a range of four orders of magnitude of roughness heights. This range included heights from sand grains to trees and buildings in cities.

It was found in this study that the roughness relationship determined by Nikuradse for sand grains is a subset of the general equation derived in this study. As a result, the derived equation may be of major importance to engineers involved with turbulent flow of liquids. For example, the derived equation should work for corrugated tubing while Nikuradse's equation works only for sand grain roughness.

Both equations for displacement height and the equation for aerodynamic roughness were verified with published data. The equation for displacement height only involves variables that are related to surface characteristics. It should be independent of atmospheric conditions. The aerodynamic roughness equation, however, involves a calibration coefficient which is related to the ratio of wind velocity at a given height to U_*. Since it is known that temperature gradients can alter this ratio, this calibration coefficient should change as atmospheric stability changes. More research is needed to evaluate how this ratio varies with temperature gradients. Obviously, experimental measurements are needed to evaluate Z_o for the stable and unstable conditions.

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