# CONTENTS

III

# Soil Conditions That Influence Wind Erosion [1]

By W. S. Chepil, *soil scientist, Soil and Water Conservation Research Division Agricultural Research Service*

## SUMMARY

Degree of cloddiness, mechanical stability of the surface crust, and bulk density and size of erodible soil fractions are some of the important primary soil factors that influence erodibility of soil by wind. Changes in the structural factors, and consequently in erodibility, are brought about by various field practices and environmental conditions.

The most erodible discrete soil particles are about 0.1 mm. in equivalent diameter. Dust tends to hinder the movement of the larger grains. The dividing point between erodible and nonerodible fractions is not distinct, for it varies with wind velocity; the equivalent size range, and the proportion of erodible and nonerodible fractions in the soil. Relatively few particles greater than 0.5 mm. in equivalent diameter are moved by common erosive winds.

Clods just large enough not to be moved by wind are most effective in preventing the movement of erodible fractions. Large clods are less effective, because in proportion to their weight they have a smaller surface with which to protect erodible particles.

The amount of erosion on a cultivated soil is limited by the height and number of clods that become exposed on the surface. At a stage when erosion ceases, the distance between the clods divided by the height of the clods remains constant for any proportion and size of nonerodible fractions present in the soil. This constant is known as the critical surface roughness constant. It has a value ranging from about 4 to 20, depending on the drag velocity of the wind and the average equivalent size of the erodible particles.

As erosion progresses, the more erodible particles are continually sorted out from the less erodible fractions. Particles moved in saltation are piled in drifts over much of the eroded area. The abrasive action of particles moved in saltation causes disintegration of the clods. The longer erosion continues, the greater is the amount of drifted material accumulated in the general vicinity of the eroded area and the lower is the subsequent velocity of the wind required to initiate erosion. There is, therefore, a range of threshold drag velocity and erodibility for any soil, depending on the previous erosional history of the affected area. Intervening rains seldom influence the threshold drag velocity and erodibility of wind-eroded fields. As soon as the soil particles on the surface are dry, erosion is resumed. Only dry soil particles are moved by wind.

Erodibility of the soil is influenced to some degree by the size, shape, and density of the structural units and to some degree by the coherence within and between these units. The former is referred to as the state of structure; the latter, as the stability of structure. The relative importance of the state and the stability of structure varies primarily with the size of the eroding field.

The resistance of different soil structural units to abrasion by impacts from windborne soil material varies directly with their mechanical stability. Stability, and hence resistance to abrasion, for the different structural units in a dry state is as follows: (1) Water-stable aggregates, (2) secondary aggregates or clods, (3) surface crust, and (4) materials among the clods cemented together and to the clods after the soil has been wetted and dried. The last of the structural units at some depth below the surface may possess mechanical stability approaching that of clods.

Since water-stable aggregates are the most stable structural units of the soil, they are the units to which the soil largely disintegrates, both by forces of the weather and by abrasive action of wind erosion. Their high stability is caused by cementing substances that are insoluble or only slightly soluble in water. The individual water-stable aggregates, or a few of them clinging together, are readily separated from the larger secondary aggregates by the wind and are usually accumulated in drifts within or near the eroded area. The fine particles are mainly carried away in the form of dust clouds, while fractions larger than the discrete water-stable aggregates remain behind as residual soil material. Dryland soils normally contain no water-stable aggregates large enough to resist movement by erosive wind. Their resistance to wind erosion consequently must depend on the formation of secondary aggregates or clods.

The identity of the clods is preserved to some degree even after repeated wetting and drying. Wetting causes some water-soluble and water-dispersible cements to become released from the originally discrete structural units and these released cements, on drying, cause a certain degree of cementation between the units. The greater the proportion in the soil of particles smaller than 0.02 mm. in diameter dispersible by water, the greater is the degree of cementation between the structural units and the greater is the resistance of the soil to breakdown by mechanical forces and abrasion from wind erosion. Also the greater the depth of soil, the greater the pressure exerted on the soil and the greater is the degree of cementation among the structural units. At considerable depth the whole soil body may become strongly cemented together. This condition is referred to as massive structure. Tillage breaks the massive structure to various sizes of blocks, referred to as clods. The clods are highly resistant to wind erosion. Implements that bring the clods to the surface without burying crop residues are most effective against erosion by wind.

The structural conditions and erodibility fluctuate in accordance with the varying influences of the seasons. In summer, increases in the proportion of the coarsest water-stable aggregates and in cloddiness are associated with increases in the proportion of the finest water-stable particles and decreases in erodibility by wind. Evidently increases in the coarsest fractions and consequent decreases in erodi-

bility are caused by increases in cementing substances contained in the finest water-dispersible fraction. In winter, the above-mentioned trends are reversed.

Approximate estimations of soil erodibility have been made from the proportion and mechanical stability of clods greater than 0.84 mm. in diameter and from the presence or absence and stability of the surface crust. Such estimations may be useful in determining the potential erodibility of different soils and soil treatments.

## INTRODUCTION

Two basic methods are used to control wind erosion. The first is to create a soil condition resistant to erosion; the second is to shelter the soil from wind. Soils differ greatly in their resistance to erosion by wind. Differences in erodibility are due to differences in their structural conditions that were either inherited or brought about by tillage, cropping, or accelerated erosion. It is important to know what soil structure would be created by different practices if wind erosion is to be reduced to the minimum.

The object of this bulletin is to bring together, analyze, and evaluate the results of research on soil conditions that influence wind erosion. References to previous publications on this and related subjects are made for readers who may be interested in the methods and procedures used in the studies and in more detailed information on specific phases of the subject.

## SCOPE OF THE SUBJECT

Wind erosion is a physical phenomenon and is therefore influenced directly by the physical conditions of the soil. Only dry soils are moved by wind (19).[2] Structure of the soil in a dry condition therefore is logically a more reliable indicator of wind erodibility than structure in a wet state.

One phase of soil structure in a dry condition is the size distribution of dry aggregates, or clods—a condition generally referred to as clod structure, or cloddiness (4, 9, 39). Cloddiness is usually determined by sieving dry soil on a nest of sieves. This technique, known as dry sieving, was used by Puchner in 1911 (34) and then by others (28, 29, 32) to characterize the soil conditions produced by tillage and cropping practices. The early methods employed sieving by hand. Later, improvements were made by substituting sieving by hand with mechanical methods (25) and by rotary instead of flat sieves (14, 20).

Resistance of the soil aggregates to breakdown by mechanical agents, such as tillage, to force of wind, and to erosional abrasion is another phase of soil structure that influences erodibility of soil by wind. The presence or absence and the condition of the surface crust also influence erodibility. Still another factor is bulk density of the erodible soil fractions. All these physical factors affect erodibility directly. They are known as the primary factors. Until the influence of the primary factors on erodibility is thoroughly understood and expressed, it will be difficult or impossible to evaluate the

---

2 Italic numbers in parentheses refer to Literature Cited, p. 38.

importance of the basic soil factors that affect the primary factors and erodibility.

Evaluation of the basic soil factors in relation to erodibility by wind falls outside of the scope of this bulletin. For want of a better name, these may be called the secondary factors. The majority of them are by no means secondary in importance. They affect erodibility by influencing the wind erosion problem. The most important of the secondary soil factors are soil texture, organic matter, soil micro-organisms and various products of organic matter decomposition, moisture, calcium carbonate, water-soluble salts, and nature of the soil colloids. Some of these factors, such as soil moisture, affect erodibility directly by affecting resistance to the forces of erosion and indirectly by influencing cloddiness and the condition of the surface crust. Moisture, therefore, may be considered as a primary or a secondary factor, depending on how it is associated with the various constituents of the soil.

Changes in structural conditions and consequently in erodibility of the soil are brought about by various field practices and environmental conditions. Some of the more important of these are climatic and seasonal conditions (17, 27, 37), kind of tillage and seeding equipment used (25, 30, 36, 38), soil moisture conditions at the time of tillage (31), kind of crops grown (24, 33), and size and layout of the fields (9, 21). It is beyond the scope of this bulletin to show how effective these practices are in influencing soil structure and erodibility, but it is important to point out that soil structural conditions and erodibility can be modified greatly by various field practices. The major objective of this bulletin is to show what soil conditions may be created to reduce erodibility of soil by wind.

## PRIMARY FACTORS THAT INFLUENCE ERODIBILITY OF SOIL BY WIND

The conditions under which wind erosion occurs are few and obvious. Wherever the surface soil is finely divided, loose, and dry; the surface is smooth and bare; and the wind is strong, erosion may be expected. By the same token, wherever the surface soil is made up of stable aggregates or clods large and dense enough to resist the force of wind; is compacted, roughened, or kept moist; or is covered by vegetation or vegetative residue; or if the wind near the ground is in any way reduced, erosion may be curtailed or eliminated. Of the six factors listed above that enhance wind erosion, four are connected directly with the condition of the surface soil. These four constitute the subject matter of this bulletin. It is important that they be thoroughly understood if they are to be properly evaluated.

### Size, Shape, and Density of Erodible Fractions

Size, shape, and bulk density of discrete soil particles considerably influence erodibility. Bulk density is defined as the weight in grams per cubic centimeter volume of a discrete soil grain or aggregate, including any air spaces within the grain or aggregate. It is con-

venient when considering erodibility by wind to express size, shape, and bulk density together by what is known as equivalent diameter. Equivalent diameter is the diameter of a standard particle that has an erodibility equal to that of a soil particle of any particular diameter, shape, and bulk density. The standard particles are spheres with a bulk density of 2.65. Graded Ottawa sand, recognized by the American Society of Testing Materials as one of its standard materials, was found to have terminal velocity of fall and erodibility much like spheres and has been used as a standard in determining the erodibility of soil grains (11). Differences in the shape of soil particles have much less influence on erodibility than their size and bulk density. In practical use, therefore, the equivalent diameter is approximately equal to $\sigma d/2.65$, in which $\sigma$ is the bulk density of the soil particles and $d$ is their diameter as determined by dry sieving.

Movement of soil particles is influenced by wind forces exerted against the surface of the ground. These forces are not dependent on velocity at some height but on the rate of increase of velocity with height, known as the *drag velocity*. For a given natural wind, the drag velocity remains the same for any surface roughness, but the velocity at all heights near the ground is influenced greatly by the surface roughness, which in turn is dependent on the overall size of the soil fractions and their arrangement on the surface. The drag velocity ($V*$), which determines the slope of the velocity distribution curve when the velocity is plotted against the logarithm of height, is equal to

$$\frac{v_z}{5.75 \log \frac{z}{k}}$$

where $v_z$ is the velocity at height $z$ and $k$ is the height at which the projected velocity curve intersects the ordinate and at which the average velocity is zero (fig. 1). Zero velocity exists somewhere among the irregularities of the surface. The greater the magnitude of surface roughness, the higher is the value of $k$ and the higher the level at which the average velocity is zero. Roughness, and hence the value of $k$, varies with size, shape, and general arrangement of the soil fractions composing the surface. The average force of wind against the ground, known as *surface drag*, can be computed from the drag velocity, since $\tau = \rho V*^2$ in which $\tau$ is the surface drag and $\rho$ is the density of the air (approximately 0.0012 in c. g. s. (centimeter-gram-second) units).

If the wind is increased gradually from a low velocity to a higher one, there comes a time when the most erodible particles are set in motion. These particles are moved along the surface of the ground in a series of jumps known as *saltation*. The higher they jump, the more energy they derive from the wind. Each time they strike the ground they transmit much of their energy to particles on the ground and cause them either to slide along the surface, move off in saltation, or be carried high in the air in true suspension. The impacts from the most erodible particles cause the movement of the larger, denser, and smaller particles. Many of the colliding particles break apart or chip away into smaller pieces. This disintegrating process is known as *abrasion*. The fragments, in turn, are moved by the wind. The eroded particles become finer as erosion progresses.
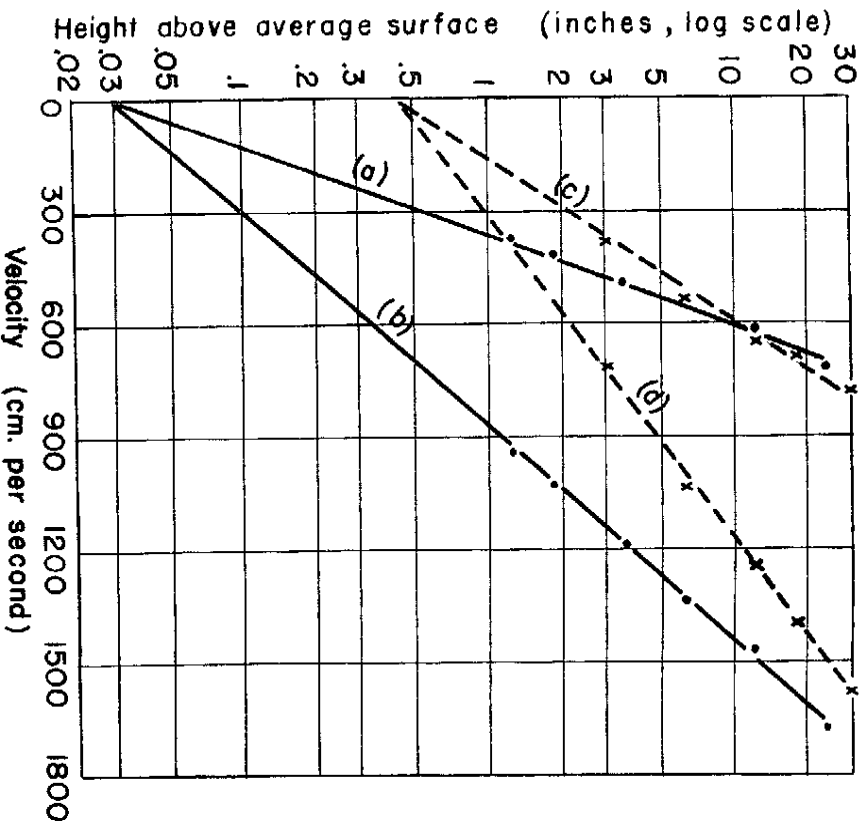
FIGURE 1.—Wind velocity distributions over wet sand and gravel surfaces of different degrees of roughness: a, Drag velocity of 41 cm. per second over a surface composed of wet sand mounds 0.16 cm. high, 1 cm. apart; b, drag velocity of 99 cm. per second over the surface as in a; c, drag velocity of 76 cm. per second over a surface composed of fine gravel mounds 5 cm. high and 30 cm. apart; d, drag velocity of 146 cm. per second over the surface as in c.

The most erodible discrete soil particles are about 0.1 mm. in equivalent diameter (fig. 2). These require a minimal drag velocity, known as the *threshold drag velocity* (designated as $V_{*t}$), of about 15 cm. per second to initiate movement. This threshold drag velocity is applicable under conditions most favorable to soil movement by wind; namely, a soil material composed only of particles 0.1 mm. in equivalent diameter, a surface that is loose, smooth, and dry, and the exposed bed is at least 30 feet long. Under those conditions the velocity required to initiate a perceptible soil movement is between 9 and 10 miles per hour at a 12-inch height. That is the lowest velocity that can produce erosion of the soil. Usually erosion does not become perceptible under field conditions until a velocity of at least 13 miles per hour at a height of 12 inches is reached. The reason for this should become more apparent as the relationship between the thresh-
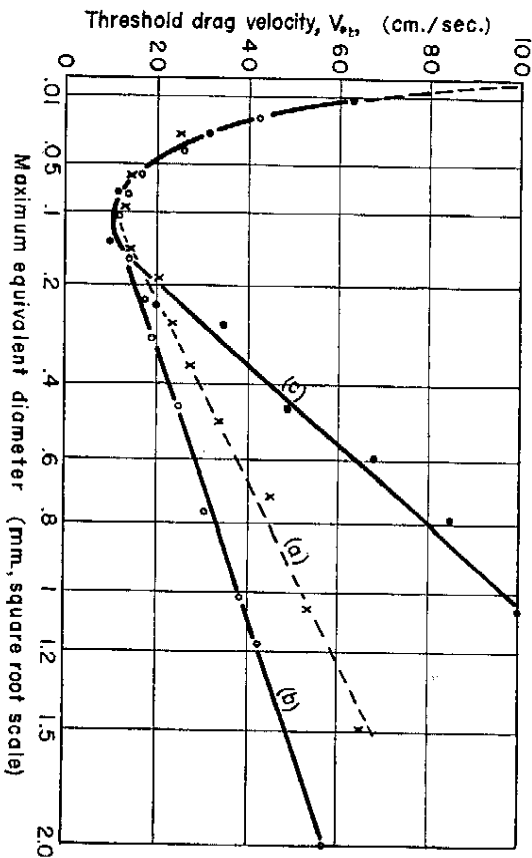
FIGURE 2.—Relation of the threshold drag velocity of the wind to the maximum equivalent diameter of the transported soil particles: a, Sieved fractions in which the ratio of minimum to maximum diameter varies as $1 : \sqrt{2}$; b, sieved fractions in which the size of particles ranges from fine dust to the indicated maximum size; c, soil containing 15 percent of nonerodible clods ranging up to 25 mm. in diameter.

old drag velocity and the equivalent size of the soil particles is explained more fully.

The threshold drag velocity increases for particles above and below 0.1 mm. in equivalent diameter (fig. 2). The threshold drag velocity for particles greater than 0.1 mm. varies as the square root of the product of equivalent diameter of the particle and the density relationship of the fluid and the particle. This square root law may be expressed by

$$V_{*t} = A\sqrt{\frac{\sigma - \rho}{\rho} gd} \qquad (1)$$

in which $d$ is the diameter of the particle, $g$ the gravity constant, $\sigma$ the density of the particle, $\rho$ is the density of the fluid, and $A$ is a coefficient whose value depends on the range of equivalent size of particles present on the eroding surface.

The relationship between the threshold velocity $v_t$ at any height $z$, equivalent diameter of the soil particles, and the roughness of the surface as exemplified by the value of $k$ can be expressed by

$$v_t = 5.75 A \sqrt{\frac{\sigma - \rho}{\rho} gd} \log \frac{z}{k} \qquad (2)$$

in which $d$ is the diameter of the particle, $\rho$ is the bulk density of the particle, and $z$, equivalent diameter of the soil particles, and the roughness of the surface as exemplified by the value of $k$ can be expressed by

As shown from equation 2, the greater the value of $k$, and the rougher the surface, the lower is the velocity (at some fixed height) required to move the particles. This relationship applies only to a condition

where the roughness elements are the soil fractions moved by the wind. It means that the larger the erodible particles or the higher they are perched on a rough surface, the higher they will protrude into the airstream and the greater the force of wind that would contribute to their movement, other factors being equal. Where the roughness elements or the surface projections or barriers are nonerodible, the threshold law expressed by equations 1 and 2 still applies but the value of coefficient A is increased considerably. Under such a condition much of the surface drag is dissipated against the nonerodible fractions and only the residual drag contributes to the movement of erodible fractions.

If the soil material is composed only of erodible fractions of a limited range of size, such as an increment of $\sqrt{2}$ commonly obtained by dry sieving, the value of coefficient A of equations 1 and 2 based on c. g. s. units is equal to about 0.1 for particles greater than 0.1 mm. in equivalent diameter (fig. 2, curve a). However, natural soil materials have a much wider range in size of fractions and therefore are associated with values of coefficient A larger and smaller than 0.1. If a soil, such as a commonly occurring dune material, is composed only of erodible fractions ranging from the largest down to the smallest erodible particles, the value of coefficient A of equations 1 and 2 is only about 0.085 (fig. 2, curve b). For such materials the threshold drag velocity varies as the square root of the average equivalent diameter of all the component particles (12). Thus, the threshold drag velocity for a mixture of different equivalent sizes of erodible particles is lower than that required to erode only the largest of the particles. Movement of the larger soil particles is facilitated by bombardment received from the smaller particles moving in saltation. The coarser fractions are transported primarily by rolling and sliding along the surface, a movement known as *surface creep.*

*Effect of Dust Particles on Soil Movement.*—Dust particles hinder the movement of the coarser grains mixed with them. The more fine dust present in the wind-eroded soil, the greater is the minimal force of wind required to initiate soil movement. The threshold velocity for these fine particles increases with the decrease in the size of particles. Loose particles smaller than 0.01 mm. if not mixed with coarser particles and if placed in a bed that is thoroughly smoothed, are not moved even by an exceedingly strong wind. For these particles the threshold drag velocity rises with the decrease in their diameter (fig. 2). No simple relationship has been found between the equivalent diameter of these fine particles and the threshold velocity required to move them.

The high resistance of the fine dust particles to movement by wind is to some degree due to cohesion among the particles. More particularly, their resistance is due to the fact that when the bed is thoroughly smoothed, the particles are too small to protrude above the viscous, nonturbulent layer of air, known as the *laminar layer,* close to the surface. It is known (26) that the soil particles of height $d$ would be submerged in the laminar layer as long as the Reynolds number of the form $V_*d/\nu$ is less than 3.5 (fig. 3). The kinematic
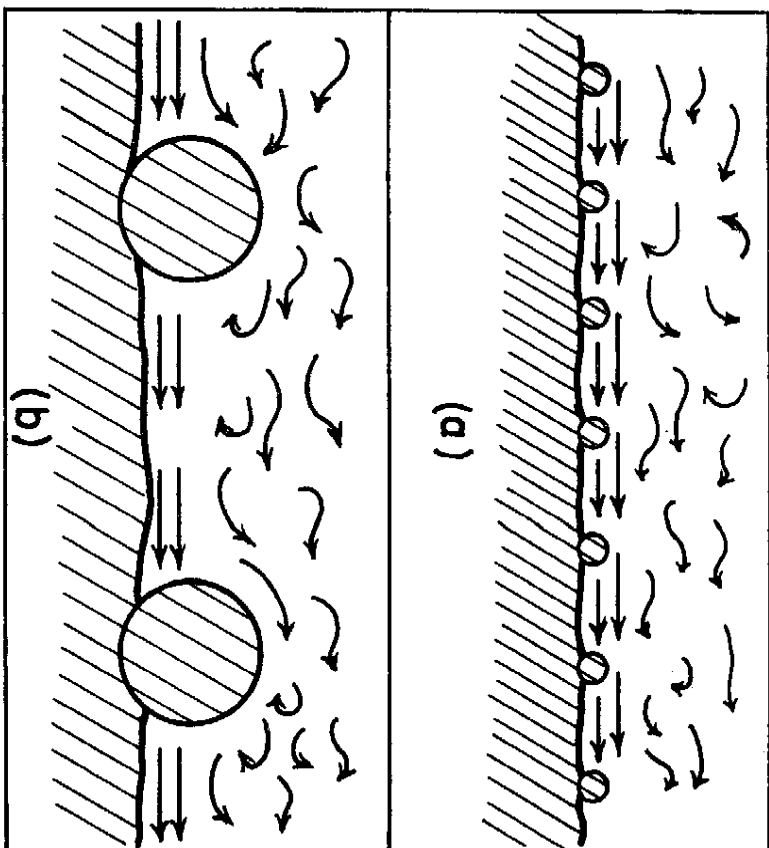


FIGURE 3.—Diagram showing (a) fine spherical particles submerged in the fluid's laminar layer (straight arrows) and (b) larger spherical grains protruding into the turbulent layer (curved arrows). Movement of soil particles is possible only in b.

viscosity, $\nu$, for air is approximately 0.15. If, on the other hand, the Reynolds number is greater than 3.5, the particles behave as obstructions in the path of the wind, throw off eddies to their lee sides, and disrupt the laminar layer. Under a force of wind equal to or greater than that required barely to move the soil particles, the particles will disrupt the laminar layer if they are greater than 0.05 mm. in diameter (6). If the surface composed of fine dust particles is roughened to a degree where the surface projections are at least 0.05 mm. in height, movement of the particles takes place under a relatively low velocity of wind. In such cases the projections composed of many dust particles clinging together are broken off and moved bodily by the wind. Movement ceases as soon as the projections are leveled down to less than 0.05 mm. in height. Under field conditions the surface roughness elements are usually much greater than 0.05 mm. The dust particles cling to the larger grains and are therefore moved readily with them.

*Rate of soil movement.*—If the wind is greater than that required barely to move the soil particles, the rate of soil movement $q$ is equal to

$$q = C\sqrt{d}\,\frac{\rho}{g}V_*^3 \qquad (3)$$

$$q = C\sqrt{d}\,\frac{\rho}{g}\left(\frac{v_z}{5.75\log\frac{z}{k}}\right)^3 \qquad (4)$$

Equations 3 and 4 show that, all other conditions remaining the same, the rate of soil movement varies directly as the cube of the drag velocity, as the square root of the average diameter of the soil particles moved by wind, and inversely as the roughness of the soil surface indicated by the value of $k$. Coefficient $C$ varies widely for different soils. It varies with the size distribution of the erodible particles (1, 4), the proportion of fine dust particles present in the mixture (4, 6), the proportion and size of nonerodible fractions (4, 10), and the amount of moisture in the soil (19). All these factors, and perhaps many more, affect the rate of soil movement and hence the value of coefficient $C$.

## Size and Total Volume of Nonerodible Fractions

On cultivated soils the nonerodible soil fractions offer a certain degree of protection to the erodible ones. For that reason the threshold drag velocity required to move the erodible particles is greater if the erodible particles are mixed with nonerodible fractions than if they alone comprise the surface soil (fig. 2, curve c). The threshold velocity law expressed by equations 1 and 2 holds just as well for mixtures of erodible and nonerodible fractions as for erodible fractions alone, but the value of coefficient $A$ is increased considerably for the mixtures. Where the nonerodible fractions comprise 15 percent of the weight of the soil, coefficient $A$ has a value of about 0.2 (fig. 2, curve c). The greater the proportion of nonerodible fractions present in the soil, the greater is the threshold drag velocity required to move a given equivalent diameter of erodible particles, and the greater is the value of coefficient $A$.

Maximum equivalent size of soil particles that can be moved by wind of a given drag velocity can be determined for each of the three distinctly different soil materials shown in figure 2. The dividing point between erodible and nonerodible fractions varies not only with the drag velocity of the wind but also with the average equivalent size, size range, and proportion of erodible and nonerodible fractions present in the soil (fig. 2). The dividing point for any wind velocity and soil condition is by no means distinct.

In all soils containing erodible and nonerodible fractions the quantity of soil removed by wind is limited by the height and number of nonerodible fractions that become exposed on the surface. If these soils are unaffected by encroachment of erodible material from the outside and if the length of the eroded area along the direction of the wind is limited, the removal of erodible fractions continues until the height of the nonerodible projections and their number per unit area

are increased to a degree that completely shelter the erodible fractions from the wind. Movement then ceases (fig. 4). The time required for movement to cease varies greatly with the soil structural conditions and the length of the field parallel to wind direction (fig. 5). The smaller the size of nonerodible fractions, the higher is the initial rate of soil movement $q$ and the shorter the time required for movement to cease. The higher the proportion of erodible to nonerodible fractions, the higher is the initial rate of soil removal and the longer the time required for movement to cease. Also, the larger the field the greater the time required for removal of erodible fractions (6, 8).

If the soil contains a large proportion of erodible fractions, few nonerodible clods per unit area of ground become exposed by the wind. The nonerodible clods under such a condition have to reach a considerable height before soil removal will cease. If, on the other hand, the soil contains a small proportion of erodible fractions, many nonerodible clods will be exposed on the surface by the wind and their height when soil movement ceases will be relatively low. The greater the number of clods exposed on the surface, the lower is their height when soil movement ceases. At a stage when soil removal ceases, the distance between the projections divided by the height of the projections remains constant for any proportion and size of nonerodible fractions present in the soil.[3] This constant is known as the *critical surface roughness constant.* It is a ratio of distance between the nonerodible surface projections to the height of the projections that will barely prevent the movement of erodible fractions by the wind. On cultivated soils this ratio has a value of 4 to 20, depending on the drag velocity and on the range and average equivalent size of the erodible fractions (9). The critical surface roughness constant of 4 means that the surface projections of height $H$ will prevent the movement of soil within a distance of $4H$ downwind of the projections. This dominant principle governing the erodibility of cultivated soils can be expressed by

$$X = KR a_1 (V_2 - V_1) \qquad (5)$$

in which $X$ is the weight of soil removable from a given area by a given wind, $V_1$ and $V_2$ are the volumes of the surface projections before and after exposure to erosive wind, respectively, $R$ is the ratio of erodible to nonerodible soil fractions, $\sigma_1$ is the bulk density of the nonerodible projections, and $K$ is a coefficient that varies with the shape, porosity, and possibly other characteristics of the projections. $V_1$ varies directly with the proportion and size of nonerodible fractions, and $V_2$ varies with the drag velocity and the size and bulk density of erodible fractions.

*Effect of Size of Field on Wind Erosion.*—The principle of surface roughness that governs the erodibility of cultivated soils is clearly manifested where the eroding area is small. The larger the area, the greater the time required for erosion to cease. In fact, in large fields removal seldom ceases for a given wind. On the average, about 120 hours of continuous exposure to erosive wind blowing from a single direction would be required to stabilize a one-half mile length. Erosive

---

[3] Distance between projections is equal to $\sqrt{\frac{1}{N}}$, where $N$ is the number of projections per unit area.
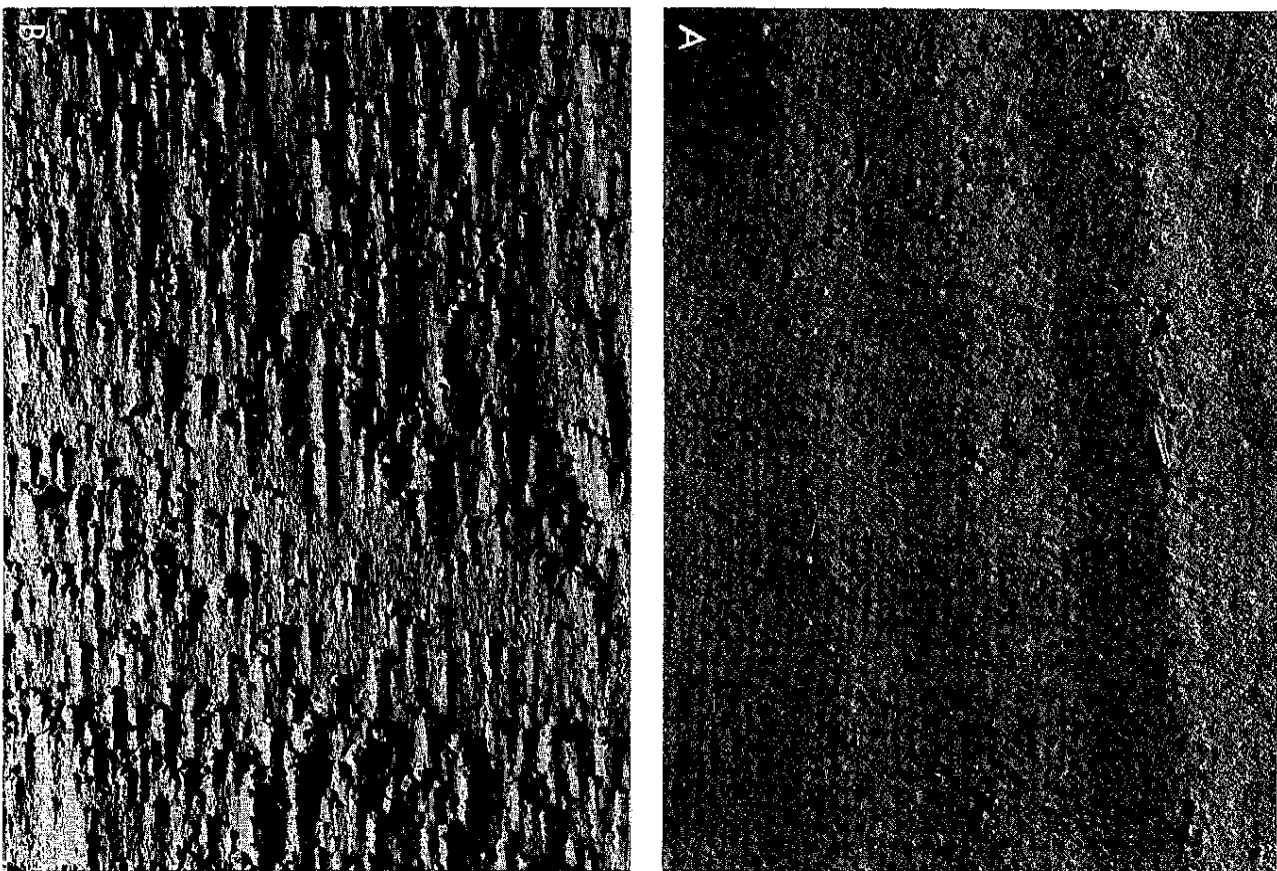
FIGURE 4.—Appearance of a silt loam composed of 92 percent erodible and 8 percent nonerodible fractions (A) before exposure to wind, and (B) after exposure for the period required for soil removal to cease. Wind velocity was 18 miles per hour at a 6-inch height and wind direction was left to right.
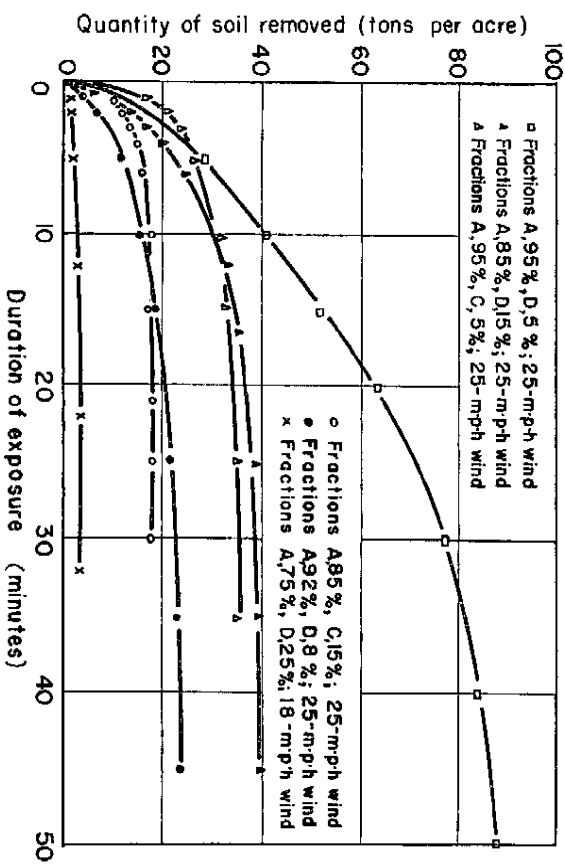
FIGURE 5.—Quantity of soil removal with duration of exposure in a wind tunnel. Soil fractions A, C, and D are less than 0.42 mm., 0.84 to 6.4 mm., and greater than 6.4 mm. in diameter, respectively. Length of soil area was 5 feet.

winds, however, seldom blow continuously from one direction for such periods. A change in wind direction also would prolong the period required to stabilize a field. Then too, great quantities of non-erodible fractions in large fields are converted to erodible particles by abrasion from the moving soil particles. The surface projections under such conditions tend to be destroyed and the rate of soil movement tends to accelerate rather than decrease, as is usual in small isolated fields. The decrease and ultimate cessation of soil movement are possible only if the surface projections or barriers are indestructible by wind erosion. The desert pavement composed of a mantle of nonerodible gravel is one example of virtual indestructibility of a stabilized surface.

*Quantity of erodible soil.*—The foregoing description of soil movement by wind has indicated that the rate of movement on cultivated soils is seldom constant but changes with the surface conditions of the soil, which, in turn, change with the duration of exposure to the wind and with the erosional history of the field. For that reason the weight of soil material removable from the surface by the wind is a more accurate measure of erodibility of dry cultivated soils than the rate of soil removal. The weight of soil material (X) that is removable from a given area by the wind may be expressed in terms of drag velocity of the wind by

$$X = CV_*^6 \qquad (6)$$

where the coefficient $C$ varies with many factors.

The quantity of erodible soil for a given drag velocity varies in great measure with the degree of soil abrasion as influenced by the characteristic length of the eroded area. For that reason it is better

to express the erodibility in dimensionless form applicable to any size of field, direction of wind, or units of measure by

$$X_1/X = CV_*^5 \tag{7}$$

in which $X$ is the weight of soil material removable from a given area under a drag velocity of 60 cm. per second, for instance; and $X_1$ is the weight removable under the same set of soil conditions under any drag velocity $V_*$.

## Soil Moisture and Rainfall Effects

Erodibility is about the same for soil that is oven-dried or air-dried. Above this range of soil moisture contents, a distinct decrease in erodibility is manifested (19). Erodibility decreases rather slowly at first, then more rapidly with increases in moisture contents, reaching zero at about the 15-atmosphere percentage for a drag velocity of about 60 cm. per second (fig. 6). The 15-atmosphere percentage is
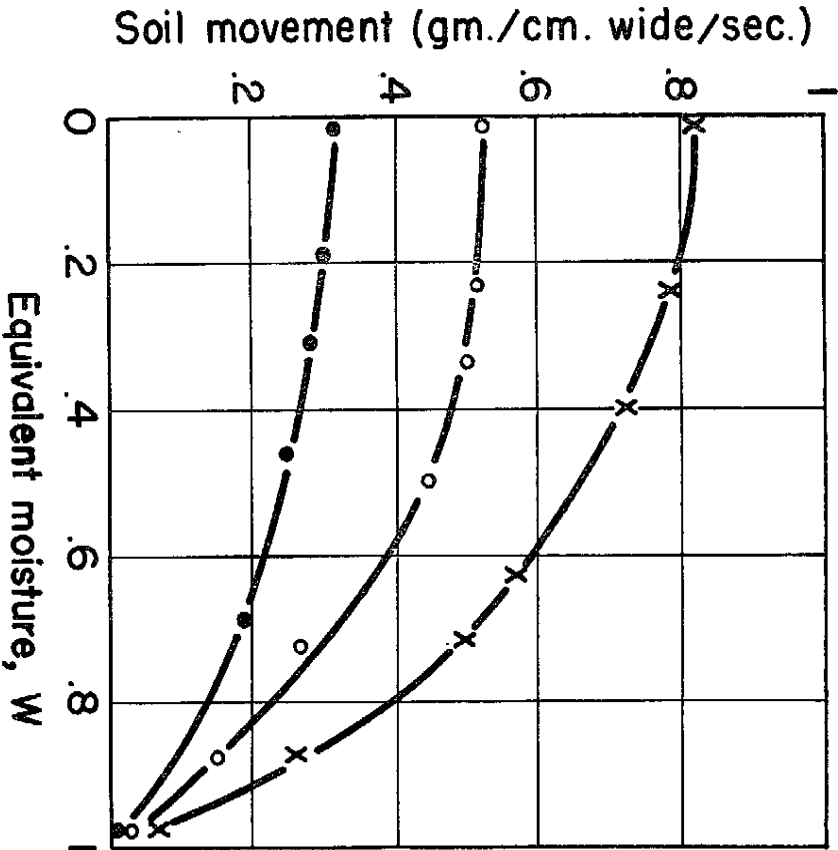


FIGURE 6.—The average influence of equivalent moisture of different soils on the rate of soil movement by wind of a drag velocity of (bottom to top) 47, 63, and 88 cm. per second.

the upper limit of hygroscopic water and corresponds approximately to percentage of water at the permanent wilting percentage. Increasing the moisture content even slightly above this limit requires a relatively great increase in the drag velocity to produce movement of the soil.

Erodibility is a function of the cohesive force of the adsorbed water films surrounding the discrete soil particles. The cohesive force among the soil particles together with the force of gravity on the particles must be overcome by the wind before erosion can occur. Therefore by utilizing equation 1, the threshold drag velocity $V_{*t}$ for moistened soil particles may be expressed by

$$V_{*t} = A\sqrt{\frac{\sigma - \rho + c}{\rho} gd} \tag{8}$$

in which $c$ is the resistance due to cohesion of the adsorbed water films exerted against lift and drag of the wind. The values of resistance $c$ were found to be equal to $6W^2$ where $W$ is the *equivalent moisture* (fig. 7). The equivalent moisture is a ratio of water content to water content at a 15-atmosphere percentage. It is equal to $w/w'$, in which $w$ is the amount of water held in the soil and $w'$ is the amount of water held by the same soil at a 15-atmosphere percentage.

Since $V_*$ is equal to $\sqrt{\frac{\tau}{\rho}}$, the rate of movement of moistened erodible particles, utilizing equation 3, may be expressed by

$$q = C_V \sqrt{g} \frac{\rho}{g} \left(\frac{\tau - c}{\rho}\right)^{1.5} \tag{9}$$

and the relative quantity of moistened soil material removable from a given area, utilizing equation 7, may be expressed by

$$X_1 X = C \left(\frac{\tau - c}{\rho}\right)^{2.5} \tag{10}$$

Equations 8, 9, and 10 apply only to conditions where moisture has been added to originally loose, dry soils. They do not apply to soils that have been moistened and then dried to various degrees, thereby causing a substantial degree of cementation of the originally discrete soil fraction—a cementation due to shrinkage of the water films on fine particles by drying.

Wetting and drying cause little cementation of drifted soil materials, such as those accumulated in drifts by wind, but they cause considerable cementation of most other soil materials. The drifted materials that cover much of the surface of eroded fields are composed essentially of water-stable grains devoid of fine dust particles required to bind them together. The impacts from a few grains moving in saltation is all that is necessary to separate the water-stable grains and to start them again in motion by the wind.

Cementation of cultivated soils by wetting and drying greatly influences erodibility. When a loose soil other than drifted material is wetted and dried, the fine particles tend to bind the whole soil body
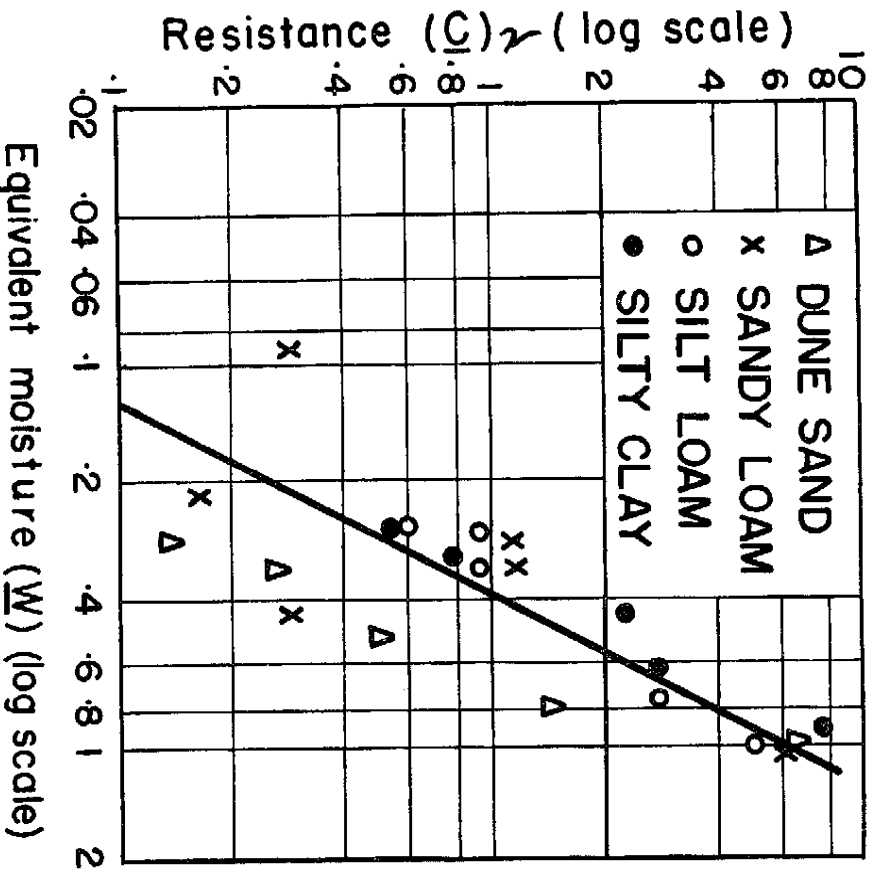
FIGURE 7.—Relation between resistance (c), due to cohesion of the adsorbed water films, and equivalent moisture (W) in various soil classes.

to form a somewhat compact mass more resistant to wind than the originally loose soil. Then, too, a surface crust is almost invariably formed, owing to impacts of raindrops on the ground. Except at the immediate surface, the primary (water-stable) aggregates and the secondary aggregates, or clods, usually undergo little transformation by individual wetting from rain and drying. A greater change occurs in the degree of compactness and cementation among the various recognizable aggregates. This type of cementation has an important influence on erodibility by wind, but the degree of cementation is generally too weak to be detectable by wet or dry sieving. Thus, wet or dry sieving, or elutriation in water or air, does not measure some important phases of soil structural stability that influence the erodibility by wind. In addition to the above-mentioned conventional methods of structural analysis, other methods must be used if erodibility is to be determined fully. One of these methods is a direct measure of stability or resistance of the various structural units to breakdown by abrasion from windborne soil particles.

## Mechanical Stability and Abradability of Soil Structural Units

Resistance of a dry soil to breakdown by mechanical agents, such as tillage, force of wind, or abrasion from windborne materials, is known as *mechanical stability*. It is due to coherence of the soil particles. Mechanical stability has been determined conveniently by dry sieving on a rotary sieve (*13*). Mechanical stability of soil fractions is equal to 100 $W_1/W$, in which $W$ is the weight of particles or aggregates greater than 0.84 mm. after the first sieving and $W_1$ is the weight after the second sieving. Mechanical stability of material among the soil fractions after they have been consolidated or cemented together by wetting and drying is equal to 100 $W_3/W_2$, in which $W_2$ is the weight of the consolidated body before sieving and $W_3$ is the weight after sieving through a sieve with openings equal to the largest of the originally discrete dry particles or aggregates. Mechanical stability of the structural units measures the relative strength of cementation or coherence *within* these units; mechanical stability of consolidated bodies of the structural units is a measure of the relative strength of cementation or coherence *among* these units.

Resistance of the various phases of field structure to disintegration by mechanical forces, such as dry sieving, is a relative measure of the resistance to disintegration by abrasion to which the soil is subjected when it is eroded by wind. Abrasion is an important phase of the wind erosion process on all soils (*7, 13*). There are two main aspects of abrasion: (1) The disintegration of nonerodible or consolidated soil units into particles small enough to be moved by wind, and (2) the wearing-away of erodible soil units to dust capable of being carried from the vicinity of the eroded region. In the first aspect, abrasion is directly associated with soil erodibility; in the second aspect, abrasion determines the mobility or the rate of removal of the fine mechanical soil constituents from the wind-eroded regions (*13*).

The relative susceptibility of the soil to abrasion by windborne soil particles has been expressed as the *coefficient of abrasion* (*18*). The coefficient of abrasion is the quantity of soil material abraded off a soil aggregate per unit weight of abrader blown against the aggregate in a 25-m. p. h. windstream. Since the amount of abrasion varies as the square of velocity, the coefficient is equal to $a\left(\dfrac{25}{v}\right)^2$, in which $a$ is the weight abraded per unit weight of abrader blown at any wind velocity ($v$) expressed in miles per hour. The coefficient of abrasion (abradability) of the different structural units of the soil varies inversely with their mechanical stability, as determined by repeated dry sieving (fig. 8). Furthermore, *modulus of rupture*, a measure of cohesive strength of soil briquets as determined by the method of Richards (*35*), varies inversely with the coefficient of abrasion and inversely with diameter of mechanical soil particles from which a briquet is formed (fig. 9).

Owing to abrasion, soil structure breaks down progressively as wind erosion continues. The amount of breakdown depends on mechanical stability of the structural units. The original initiation of perceptible soil movement for the first time in the field generally requires a much higher drag velocity than for succeeding windstorms; the soil is usually
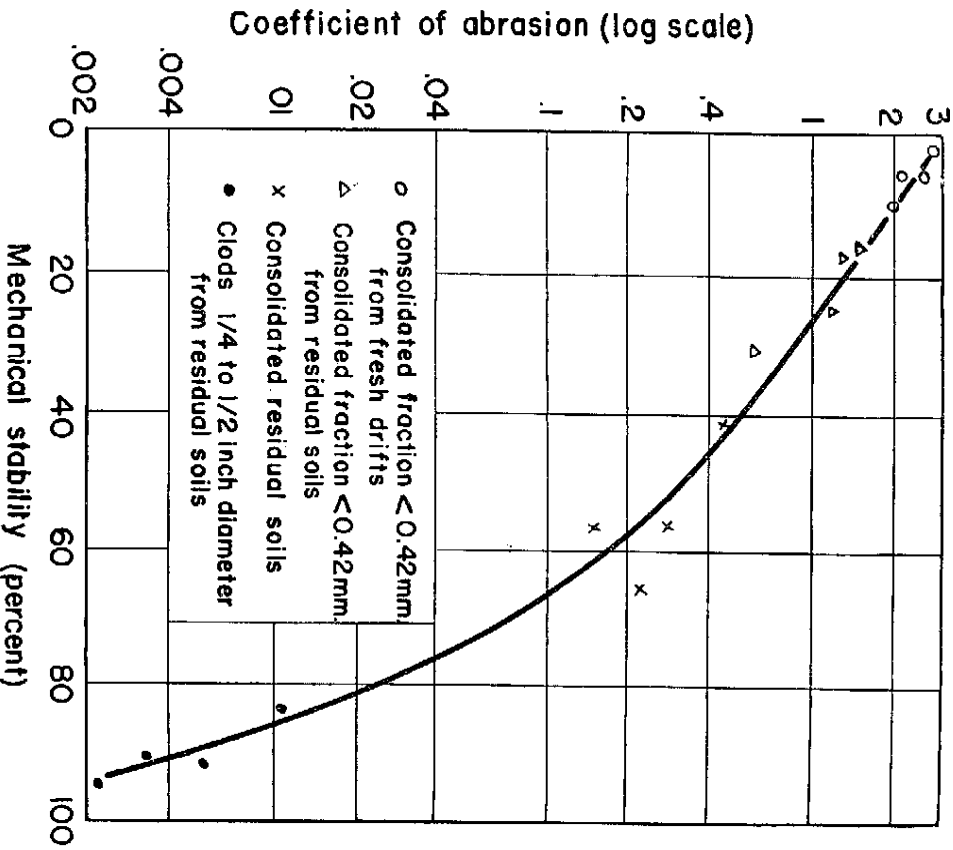
FIGURE 8.—Relation between the coefficient of abrasion and mechanical stability of different phases of field structure of soil.

covered with a thin surface crust that is somewhat resistant to wind erosion. As soon as some soil particles are loosened and moved by wind, their abrasive action against the surface causes the crust to disintegrate and expose a more highly erodible soil beneath. Then, too, the nonerodible clods gradually become broken down by impacts of saltating grains. The erodible fractions are being continually sorted from the less erodible fractions and usually are piled in hummocks in the vicinity of the eroded area. The longer erosion continues, the greater is the quantity of highly erodible material accumulated on the leeward side of an isolated field and the lower is the velocity of wind required to initiate erosion of the field. Soil movement usually begins and is of greatest intensity on the leeward side of the field where the concentration of the eroding particles is the
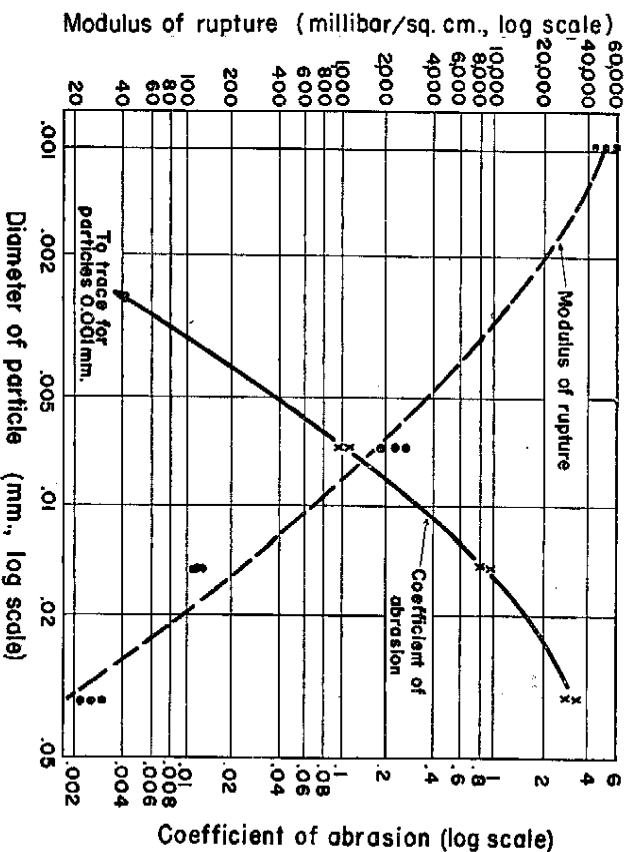
FIGURE 9.—Relation of modulus of rupture and coefficient of abrasion to diameter of mechanical soil particles in dry briquets.

greatest and abrasion of the crust and of the nonerodible soil fractions is most intense (8).

Therefore, a range of threshold drag velocity for any soil depends on the previous erosional history of the field. This range varies from the original threshold velocity of the previously noneroded field to the threshold velocity of dry dune materials. This range is between 13 and 30 m. p. h. at 1-foot height (6). Once a field has been exposed to a series of erosive winds, it generally starts to erode when the wind reaches a velocity of about 13 m. p. h. at 1-foot height—a velocity sufficient to move dry dune materials. This threshold velocity is remarkably uniform for all dune materials.

Surface soil, such as exists in the field after wetting and drying, is not homogeneous, although often it appears to be so. It is composed of various types of structural units cemented together in varying degrees (16). The strength of cementation and, consequently, the abradability when the soil is dry vary greatly for different soils and different structural units of the soil. Two types of soil cements seem to be responsible for consolidation of the soil in different structural units: (1) Water-insoluble; and (2) water-soluble or water-dispersible. These cements appear to be responsible for the following types of structural units with distinct degrees of mechanical stability and abradability by wind: (1) Water-stable aggregates; (2) secondary aggregates, or clods; (3) fine materials among the secondary aggregates; and (4) the surface crust. Those phases of field structure in cultivated soils are shown in figure 10. Each secondary aggregate in figure 10
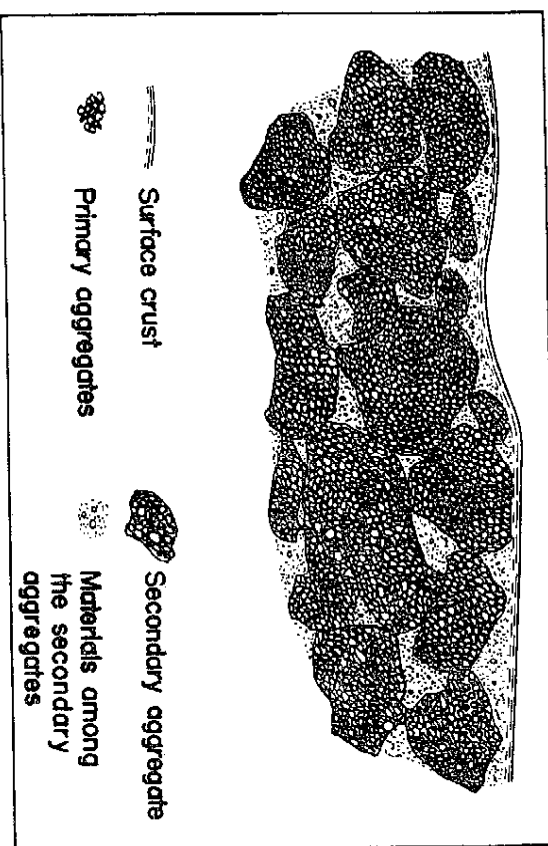
is designated by a line surrounding a number of primary aggregates, of which the secondary aggregate is composed.

*Water-stable aggregates.*—These primary aggregates, which seldom exceed 1 mm. in diameter in cultivated dryland soils, are held together by water-insoluble cements composed of clay particles and irreversible or slowly reversible inorganic and organic colloids (*9*). The water-stable granules possess high strength of coherence (*13, 16*) and stability against the disintegrating forces of the weather (*17*). Since they are the most stable structural units of the soil, they represent the units to which the secondary aggregates ultimately disintegrate, both by forces of weather and by abrasive action of wind-eroded soil particles. The water-stable aggregates are readily separated from the other soil fractions by the wind and are usually accumulated in drifts or mounds within and outside the eroded fields. Particles finer than water-stable aggregates are removed in the form of dust, while the coarser fractions (clods, gravel, and rocks) remain behind as residual soil materials.

The drifted particles are principally individual water-stable aggregates or discrete sand grains (table 1). The drifted sand grains and clay aggregates exhibit the greatest mechanical stability, while those of intermediate texture exhibit a somewhat lower mechanical stability (table 2). Without appreciable quantities of fine dust, the wind-blown grains tend to remain as discrete units, giving the soil materials a characteristically mellow structure commonly referred to as "good tilth." Dryland soils are virtually devoid of water-stable particles large enough to resist movement by wind. Their resistance to wind erosion consequently must depend on the formation of secondary aggregates or clods.

FIGURE 10.—Diagrammatic representation of structure of cultivated soil after wetting by rain and drying.



Surface crust

Primary aggregates

Secondary aggregate

Materials among the secondary aggregates

TABLE 1.—*Size distribution of dry and water-stable aggregates in wind-eroded (drifted) and residual soil materials*

| Soil textural class | Soil material | Dry fractions | | | Water-stable fractions | | |
|---|---|---|---|---|---|---|---|
| | | >0.84 mm. | 0.84–0.05 mm. | <0.05 mm. | >0.84 mm. | 0.84–0.02 mm. | <0.02 mm. |
| | | *Percent* | *Percent* | *Percent* | *Percent* | *Percent* | *Percent* |
| Sand | Drifted | 1.1 | 98.0 | 0.9 | 1.0 | 97.8 | 1.2 |
| | Residual | 18.5 | 79.9 | 1.6 | .8 | 97.4 | 1.8 |
| Loamy sand | Drifted | 2.3 | 96.7 | 1.0 | 4.3 | 93.9 | 1.8 |
| | Residual | 27.2 | 69.6 | 3.2 | 2.5 | 93.7 | 3.8 |
| Sandy loam | Drifted | 5.4 | 92.5 | 2.1 | 3.0 | 94.4 | 2.6 |
| | Residual | 27.8 | 67.6 | 4.6 | 1.4 | 93.3 | 5.3 |
| Loam | Drifted | 3.1 | 90.3 | 6.6 | 1.9 | 93.2 | 4.9 |
| | Residual | 39.4 | 51.0 | 9.6 | 4.1 | 86.2 | 9.7 |
| Silt loam and silty clay loam | Drifted | 9.6 | 86.3 | 4.1 | 4.4 | 87.4 | 8.2 |
| | Residual | 37.8 | 53.6 | 8.6 | 5.0 | 82.3 | 12.7 |
| Silty clay and clay | Drifted | 24.7 | 74.5 | .8 | 9.6 | 86.2 | 4.2 |
| | Residual | 28.4 | 70.5 | 1.1 | 4.8 | 90.8 | 4.4 |

TABLE 2.—*Mechanical stability of different structural units and of fine materials among the structural units of wind-eroded and residual soil materials*

| Structural units | Mechanical stability | | | |
|---|---|---|---|---|
| | Sandy loam | Silt loam | Silty clay loam | Clay |
| | *Percent* | *Percent* | *Percent* | *Percent* |
| Particles >0.42 mm., from fresh drifts (chiefly water-stable) | [1] 97.6 | 95.5 | 95.0 | 97.0 |
| Dry aggregates or clods >0.42 mm., obtained by dry sieving | 83.8 | 91.7 | 90.6 | 93.8 |
| Surface crust ⅛- to ¾-inch thick on residual soil | 60.2 | 73.3 | 69.3 | 58.5 |
| Particles <0.42 mm. from residual soils after consolidation [2] | 17.0 | 28.1 | 27.3 | 17.4 |
| Particles <0.42 mm. from fresh drifts after consolidation [2] | [3] 3.0 | 8.8 | 5.0 | 4.6 |

[1] Mostly sand grains.
[2] Consolidation was accomplished by spraying the dry soil material in a column 2 inches high with 1 inch of water followed by drying.
[3] Cementing strength among particles was barely overcome by wind having a drag velocity of about 60 cm. per second.

*Secondary aggregates or clods.*—Secondary aggregates are next in order of mechanical stability, depending on soil class, depth, and tillage treatment. They are held together in a dry state primarily by water-dispersible cements acting under pressure from depth and time. The cements are composed mainly of water-dispersible particles smaller than 0.02 mm. in diameter (table 3). When these fine particles are removed by repeated decantation after shaking in water, the water-stable aggregates to which the clods disintegrate after shaking in water are much like sand grains in that they fail to cohere to each other after a layer of them is dried (table 3). Fine water-dispersible particles are necessary to bind the water-stable aggregates together to form clods.

The clods are resistant to wind erosion so long as they remain large enough to resist movement by wind. Many of them maintain their identity for some time after repeated wetting and drying in the field. Individual rains have little influence on the form or compactness of clods below the surface even after they lose their visible identity after the soil is wetted and dried. Only within a narrow zone of the immediate surface where the soil mass assumes a structure distinctly different from that below do the clods become appreciably disintegrated by impacts of raindrops. Abrasive tests have indicated that after repeated wetting and drying the clods become merely embedded in the fine, loosely consolidated portion of the soil. The strength of cementation between the clods is generally much lower than within the clods; hence, the reason why blocks of soil abrade unevenly when exposed to impacts of windborne soil grains (fig. 11).

TABLE 3.—*Relation between dry clod formation and percentage of particles <0.02 mm. dispersed in water*

| Soil material and treatment | Particles <0.02 mm. dispersed in water | Clods >0.42 mm. after dry sieving |
|---|---|---|
| Soil textural class | | |
| | *Percent* | *Percent* |
| Dry sieve fraction<0.42 mm., consolidated.[1] | | |
| Sandy loam | 10.2 | 17.0 |
| Silt loam | 19.3 | 28.1 |
| Silty clay loam | 18.2 | 27.3 |
| Clay | 9.8 | 17.4 |
| Dry sieve, fraction<0.42 mm., from which particles<0.05 mm. were removed by shaking and repeated decantation in water, and then consolidated.[1] | | |
| Sandy loam | 0 | 0 |
| Silt loam | 0 | 0 |
| Silty clay loam | 0 | .09 |
| Clay | 0 | .23 |

[1] Consolidation was accomplished by spraying dry soil material in a column 2 inches high with 1 inch of water followed by drying.

*Materials among the clods.*—The cohesive forces that exist among the clods after the soil has been wetted and dried vary greatly, as within the clods, depending on the number and the nature of wettings, on the depth and consequent pressure exerted against the soil, and on the physical-chemical nature of the soil. The degree of cementation that holds the clods together after the soil has been wetted and dried is due in large measure to the quantity of particles of the size of silt and clay dispersible in water (table 4). Wetting apparently causes either some water-soluble and water-dispersible cements or water-dispersible cements to become released from the originally discrete structural units and, on drying, the cement causes a certain degree of cementation between the units. The greater the quantity of fine particles dispersible by water, the greater is the degree of cementation among the structural units and the greater is the resistance of the soil to breakdown by mechanical forces.

Pressure likewise increases the cementation among the clods and other structural units. The greater the depth, the greater is the pressure exerted on the soil and the greater is the degree of cementation and mechanical stability among the structural units, until the whole soil mass, at a certain depth, may become strongly cemented together. This condition is often referred to as a massive structure. Tillage breaks the structure to various sizes of blocks referred to as clods. Tillage, if suitable, may bring the clods to the surface to resist erosion by wind. But it also tends to bury the clods to the crop residue. Implements that perform effectively the dual purpose of increasing the surface cloddiness and, at the same time, avoiding the burial of crop residues are needed.
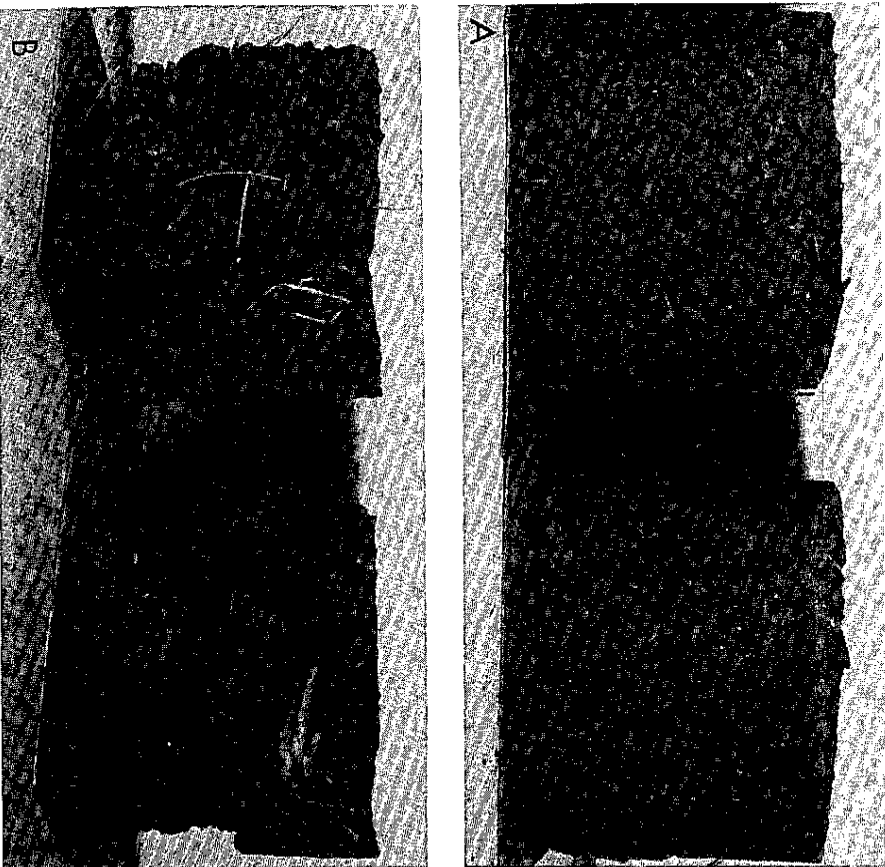
FIGURE 11.—Blocks of silt loam (left) and sandy loam (right) consolidated by spraying with 1 inch of water 4 times and drying after each wetting: A, Before abrasion by dune sand; B, after abrasion. The originally embedded, less abradable soil aggregates are exposed after abrasion. They are mainly secondary aggregates, or clods.

The fine particles that tend to cement the clods and other structural units together are composed of silt, clay, and various materials of organic and inorganic origin. Dispersed silt, although usually not considered as a soil cement, acts as a weak cement of sufficient strength to resist considerably the force of wind (18). Silt particles are dispersed by water much more readily than particles of clay size. The presence of large quantities of dispersed silt particles in a soil appears to cause the formation of a compact, massive structure, which, while quite resistant to wind erosion, may present a serious structural problem otherwise. Bradfield and Jamison (3) concluded that hard and intractible soils were usually those largely composed of fine silt having a single-grain structure when dispersed in water.

TABLE 4.—*Relation between mechanical stability of blocks of consolidated soil and the percentage of particles <0.02 mm. dispersed in water*

| Soil material and treatment | Soil textural class | Particles <0.02 mm. dispersed in water | Mechanical stability [1] |
|---|---|---|---|
| | | Percent | Percent |
| Dry-sieve fraction <0.42 mm. from residual soil, consolidated. | Sandy loam | 10.2 | 17.0 |
| | Silt loam | 19.2 | 28.1 |
| | Silty clay loam | 18.2 | 27.3 |
| | Clay | 9.8 | 17.4 |
| Dry-sieve fraction <0.42 mm. from drifts, consolidated. | Sandy loam | 3.9 | 3.0 |
| | Silt loam | 8.8 | 8.8 |
| | Silty clay loam | 7.2 | 5.0 |
| | Clay | 3.0 | 4.6 |

[1] Percentage of clods >0.42 mm. after dry sieving the blocks on a 0.42-mm. rotary sieve.

*The surface crust.*—Because of impacts of rain, the soil material at the surface becomes more dispersed than the soil below. On drying, the dispersed soil forms a thin surface crust that is more compact and mechanically stable than some parts of the soil below. The crust often does not exceed one-sixteenth inch in thickness, but occasionally it may reach a thickness of one-fourth inch. The crust is easily recognizable by its dense, platy structure. This type of structure becomes less distinct with depth, until it merges with the soil below. Medium-textured soils containing a high proportion of silt are most subject to dispersion in water and, therefore, these soils produce the thickest and most compact crust (table 5). That condition contributes to the usually high resistance of the medium-textured soils to erosion by wind. Sandy soils generally are less subject to surface crust formation, because they do not contain a high proportion of silt and clay. That property contributes considerably to the high erodibility of sandy soils by wind. Clay soils are highly variable with respect to wind erosion. Those that contain a high proportion of fine water-dispersible particles tend to puddle and resist erosion by wind. On the other hand, some clays are not subject to a high degree of dispersion (table 5); consequently, the surface crust and the clods tend to remain as fine granules, some of which are readily moved by wind.

A rain or a series of rains often carries some of the finely dispersed and water-soluble cementing materials downward, leaving the coarser particles, such as sand or water-stable aggregates, at the top. Some of these coarser particles remain loose on the surface and often contribute to the initial stage of wind erosion. Being on the surface, they dry rapidly. Consequently these coarser particles may be moved by wind soon after a rain, even before the drying of the surface has become apparent. Abrasion from these particles tends to wear down the surface crust, to hasten the drying of the surface, and to accelerate the soil movement as long as the wind that is strong enough to move the soil material continues. Small showers often tend to

smooth the soil surface, to loosen some of the surface particles, and, if the field is large, to accelerate rather than alleviate soil movement by wind.

TABLE 5.—*Relation between mechanical stability of the surface crust and percentage of particles <0.02 mm. dispersed in water*

| Soil textural class | Soil material | Particles <0.02 mm. dispersed in water | Mechanical stability of crust |
|---|---|---|---|
| | | *Percent* | *Percent* |
| Sandy loam | Drifted | 6. 2 | 44. 7 |
| | Residual | 10. 4 | 60. 2 |
| Silt loam | Drifted | 10. 2 | 60. 8 |
| | Residual | 16. 8 | 73. 2 |
| Silty clay loam | Drifted | 10. 4 | 59. 7 |
| | Residual | 15. 4 | 69. 3 |
| Clay | Drifted | 4. 9 | 38. 1 |
| | Residual | 9. 6 | 58. 5 |

On many soils the rate of soil movement is slow at the beginning, but it accelerates as the surface crust is worn through and a weakly consolidated soil beneath it is exposed to the wind (*13, 16*). The nature of the surface crust and its relation to erosion by wind perhaps can be interpreted best from its appearance as it is destroyed by abrasion with dune sand (fig. 12). The surface crust was completely



FIGURE 12.—Surface crust on clay soil partly destroyed by abrasion with dune sand for 5 minutes with a wind velocity of 28 miles per hour at 12-inch height.

stable under the same wind of 28 miles per hour without the abrader.

*Order of mechanical stability.*—Susceptibility of the soil to abrasion by wind varies inversely with its mechanical stability (fig. 8). The order of mechanical stability from highest to lowest, and hence the order of abradability from lowest to highest, for the different structural units in a dry state are as follows: (1) Water-stable aggregates, (2) secondary aggregates or clods, (3) surface crust, and (4) fine materials among the clods cemented together and to the clods after the soil has been wetted and dried. The last of the structural units at some depth below the surface may possess mechanical stability approaching that of clods.

Mechanical stability tends to reduce wind erosion by resisting the breakdown of nonerodible units to smaller erodible particles. The breakdown in the field is caused by two groups of commonly occurring agents: (1) Mechanical agents such as tillage machinery, and (2) abrasive action of windborne soil material.

## RELATIVE INFLUENCE OF STATE OF STRUCTURE AND OF STABILITY OF STRUCTURE ON ERODIBILITY BY WIND

Erodibility of the soil is dependent (1) to some degree on size, shape, and density of the structural units, and (2) to some degree on the mechanical stability of the structural units. The first may be referred to as the state of structure and the latter as the stability of structure. Both phases of structure are measurable by elutriation, dry sieving, and repeated dry sieving (*13*). The relative importance of the state and stability of structure with respect to erodibility by wind varies with the area of the field, the roughness of the surface, and many other factors. If the area of the field is small, the amount of abrasion from erosion is small and erodibility of the field is determined primarily by the state of structure, or specifically by the proportion of discrete particles small enough to be moved by wind. If on the other hand the field is large, mechanical stability of the structural units is the more important factor. In such case, if the soil structural unit lacks mechanical stability, the presence of even a small quantity of loose, erodible material on the surface is usually sufficient for substantial disintegration of the structural units by abrasion from windborne material and for consequent intense erosion of the loosely cemented soil (*13*).

The relative importance of the state and stability of structure of different soils is shown in table 6, based on wind-tunnel tests. A surface crust formed by spraying the soil with water followed by drying (condition b) reduced greatly the quantity of soil material eroded by wind. However, when the soil was subjected to impacts of soil particles blown in from the outside (condition c), the crust soon was worn through and the rate of soil removal was increased considerably and continued as long as the stream of sand passed over the soil. The amounts of erosion occurring under condition b are comparable to those obtained in small, isolated fields where abrasion is limited; the amounts of erosion occurring under condition c, on the other hand, are

applicable to those on the leeward sides of large, open fields where the intensity of abrasion from eroded particles is relatively great.

TABLE 6.—*The influence of state of structure and stability of soil structure on erodibility by wind*

| Soil class | Clods >0.5 mm. equivalent diameter | Degree of cementation between the clods after consolidation | Amount or rate of soil erosion [1] — | | |
|---|---|---|---|---|---|
| | | | Condition a | Condition b | Condition c |
| | *Percent* | *Percent* | *Tons per acre* [2] | *Tons per acre* [2] | *Tons per acre per minute* |
| Sandy loam | 39.8 | 17.0 | 3.4 | 0.4 | 13.0 |
| Silt loam | 32.3 | 28.1 | 4.5 | .2 | 5.6 |
| Silty clay loam | 42.1 | 27.3 | 2.9 | .3 | 9.4 |
| Clay | 12.1 | 17.4 | 9.5 | 3.4 | 11.0 |

[1] Conditions:
a—Exposure to wind of well-mixed, loose, and dry soils.
b—Exposure to wind after consolidating the soil by spraying with 1 inch of water and drying.
c—Exposure to wind and a stream of windborne sand after consolidating the soil. Rate of sand flow was 1,000 grams per minute per 8-inch width.
[2] Until movement ceased.

# SEASONAL VARIATIONS IN CLODDINESS, MECHANICAL STABILITY, AND ERODIBILITY BY WIND

Biological activities and alternating wetting and drying and freezing and thawing appear to have a strong influence on the structural conditions and erodibility of soil by wind. The structural conditions and erodibility fluctuate in accordance with the varying influences of the seasons.

Soil cloddiness and mechanical stability of clods are decreased and erodibility increased in winter in cases where the soil is moistened at least occasionally (fig. 13). Also, the changes are greatest at or near the surface of the ground and least, if any, at a 6-inch depth (table 7). A visible change in cloddiness of moist soils from fall to spring is shown in figure 14.

Irrespective of the seasonal variations in structure and erodibility of soil at and near the surface of the ground, the degree of cloddiness and mechanical stability of clods increases and erodibility by wind decreases with depth in all soils (figs. 15 and 16). Cloddiness and mechanical stability of clods also increase and erodibility decreases with the fineness of soil texture; that is, a soil with percentage of
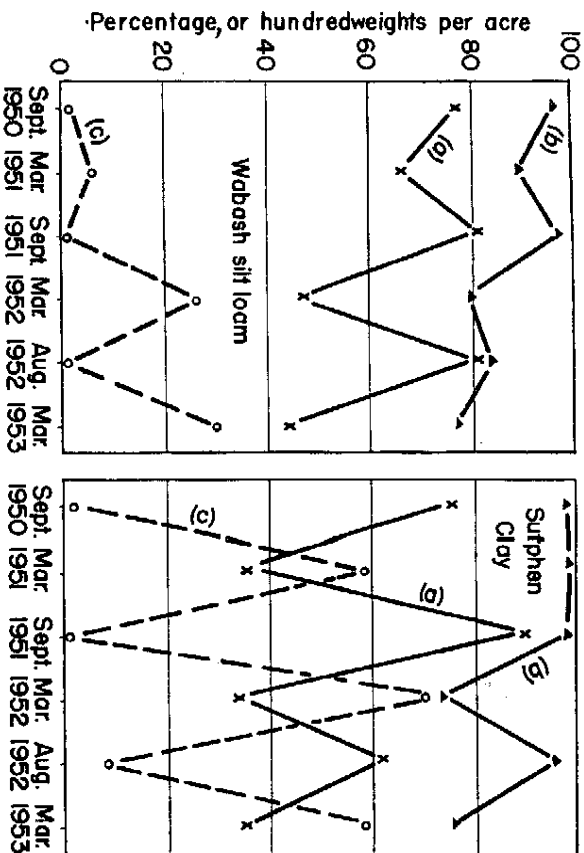
FIGURE 13.—Seasonal fluctuation in dry soil structure and erodibility by wind: *a*, Percentage of dry clods >0.84 mm. in diameter; *b*, percentage of mechanical stability of dry clods; and *c*, erodibility in hundredweights per acre. All measurements were based on soil from surface to 1-inch depth.

clay up to 20 to 28 percent, depending on the nature of the clay (*15*, *18*). Increases in clay beyond these percentages decrease cloddiness and degree of cementation between the clods, increase the erodibility by wind, but continue to increase the mechanical stability of the clods.

TABLE 7.—*Influence of seasons on some phases of soil structure and erodibility at various depths*

[Averages for Cass loam during a 3-year period at Manhattan, Kans.]

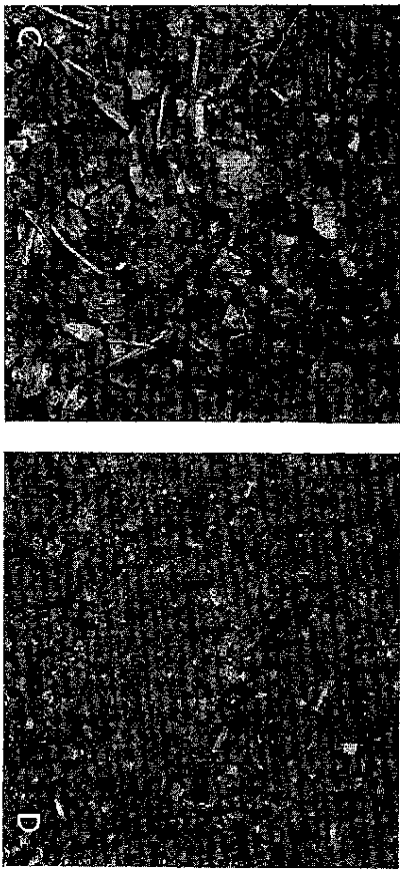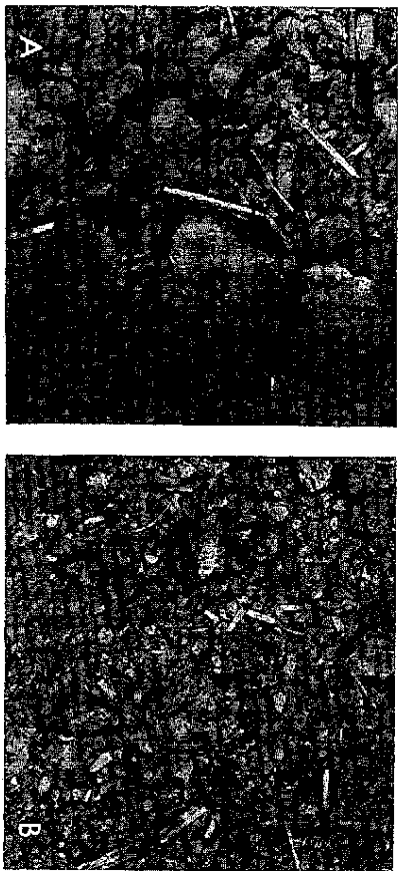| Depth (inches) | Season | Clods >0.84 mm. | Mechanical stability of clods | Amount eroded in tunnel until movement ceased |
|---|---|---|---|---|
| | | *Percent* | *Percent* | *Tons per acre* |
| 0 to 1 | Fall | 65.0 | 87.8 | 0.4 |
| | Spring | 46.7 | 72.7 | 1.5 |
| 1 to 3 | Fall | 71.9 | 87.8 | .24 |
| | Spring | 58.1 | 80.0 | .8 |
| 3 to 6 | Fall | 80.5 | 88.8 | .06 |
| | Spring | 80.5 | 90.6 | .09 |

FIGURE 14.—Appearance of soil from surface to 1-inch depth after sampling, drying, and thorough mixing: A, Wabash silt loam in fall 1951; B, Wabash silt loam in spring 1952; C, Sutphen clay in fall 1951; D, Sutphen clay in spring 1952.



Clods >0.84 mm. (percent)

- Loamy sand
- x Sandy loam
- o Loam, clay loam, sandy clay loam

Erodibility (tons per acre, log scale)

Depth (inches)
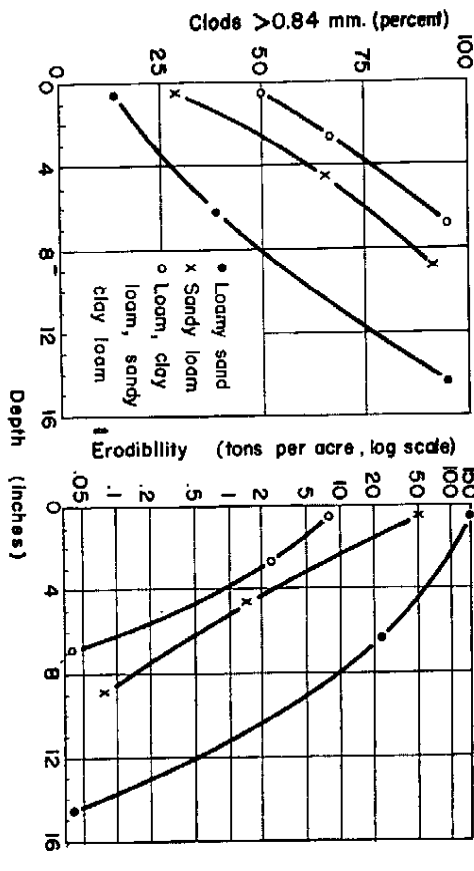
FIGURE 15.—Proportion of clods and erodibility at various depths for different soil classes.

Mechanical stability of clods (percent)

Depth (inches)

- Loamy sand
- x Sandy loam
- o Sandy clay loam, Clay loam, Loam

FIGURE 16.—Mechanical stability of clods of various soil classes at different depths.

Increased cloddiness and mechanical stability of clods with depth are due partly to an increase in the fineness of soil texture and partly to degree of soil compaction. Some types of tillage tend to bring up cloddy soil from lower depths and thereby reduce erodibility by wind.

The effects of tillage are temporary, because the forces of the weather, especially freezing and thawing of moist soil during the winter, tend to break the clods to sizes small enough to be moved by 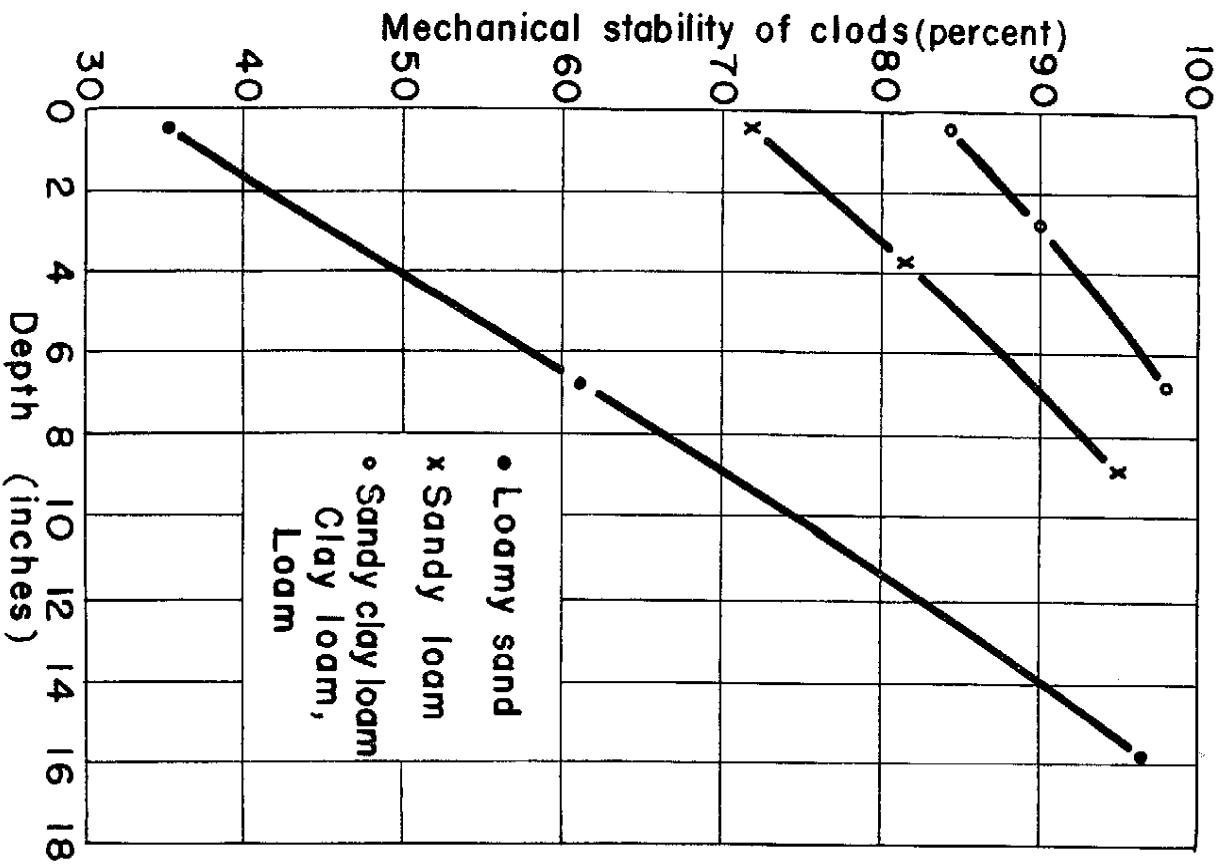wind. But as the clods at the surface are broken down, clods below the surface are being formed. Hence, repeated tillage of a proper type is useful in maintaining a cloddy surface indefinitely. The degree of cloddiness that can be maintained varies with the nature of the soil and with the depth and nature of tillage.

## ESTIMATING ERODIBILITY BY WIND

The description of the relationships between the various soil structural factors and erodibility by wind indicates generally what constitutes an erodible and a nonerodible soil. Attempts have been made to estimate soil erodibility from these relationships (4, 12, 22). Factors recognized in the estimates were the proportion of erodible fractions in relation to the drag velocity of the wind, the volume of nonerodible fractions, and the mean weighted equivalent diameter of the erodible units. They are by no means all the factors that influence erodibility. The relationship between the various phases of soil structure and erodibility is complicated and varied. However, a method of estimating the relative erodibility must be reasonably simple if it is to be practical. The two most important criteria of resistance of soil to erosion by wind are soil cloddiness and mechanical stability of clods and surface crust.

The dividing line between erodible and nonerodible fractions for mineral soils is about 0.84 mm., one of the sizes of square sieve openings in a sieve series proposed in 1919 by the United States Bureau of Standards. A curve based on wind-tunnel tests expressing an average relationship between the quantity of soil eroded when dry and the proportion of clods greater than 0.84 mm. as determined by dry sieving, is shown in figure 17. This figure is based on two groups of measurements reported previously (13, 23). The quantities of erosion are based on (1) a soil surface leveled by hand over which the roughness varies somewhat, depending on the size of the soil aggregates; (2) a soil that is loose, uniformly mixed, and free from organic residues; (3) a soil surface that is thoroughly air-dried; (4) a 5-foot length of the exposed soil area; (5) a drag velocity of 61 cm. per second; and (6) a wind free from gusts and blowing from one direction. A change in any of the listed conditions would have produced a change in the quantity of eroded soil. The quantities of eroded soil indicate the quantities removable before movement ceased. They indicate the quantities removable under some definite wind blowing from one direction. Because of the short length of the exposed area, abrasion by impacts from saltation that commonly occur in the field was almost absent. The quantities of soil erodible in the wind tunnel may be expected, therefore, to be substantially lower than the quantities in the open field. Nevertheless, the basis that determines the relative degree of erosion from field areas and from small areas in the tunnel is apparently the same. This basis is the quantity of erodible fractions removable from the surface of the soil by the wind.

Quantity eroded (X)   (tons/acre, log scale)

Legend:
- Sand and loamy sand
- Sandy loam
- Loam, silt loam, clay loam and silty clay loam
- Clay and silty clay

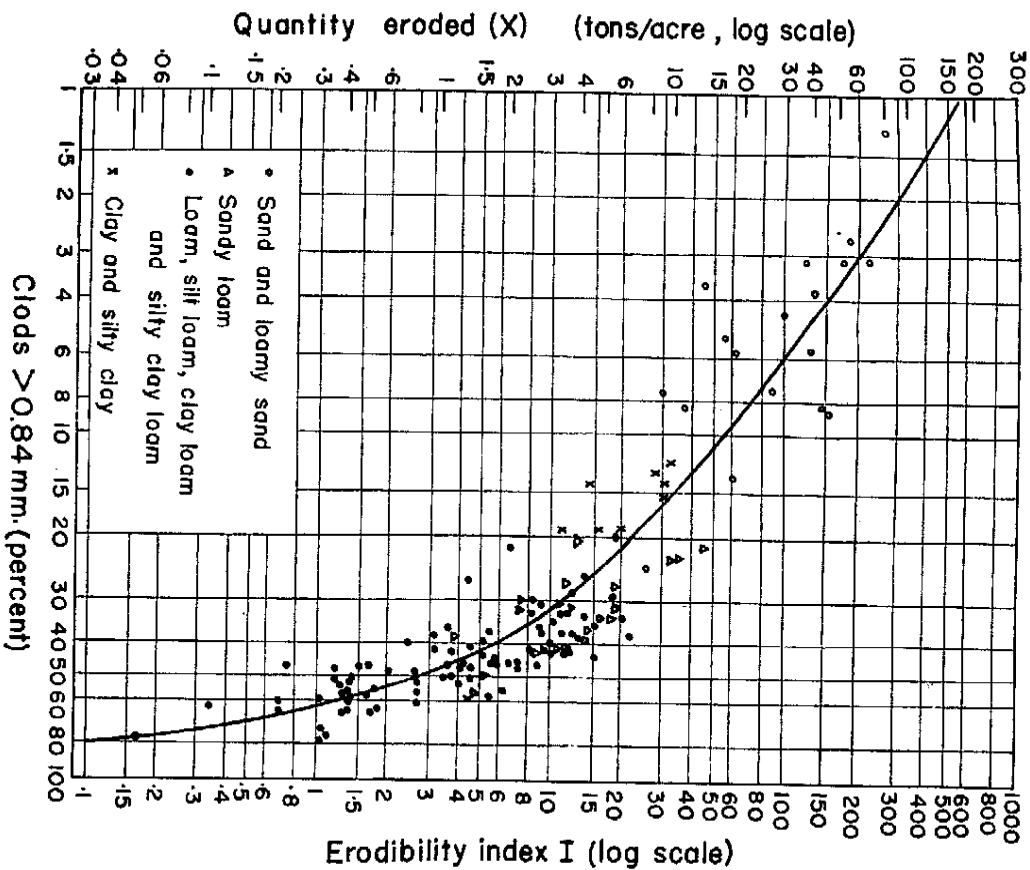Clods >0.84 mm. (percent)

Erodibility index I (log scale)

FIGURE 17.—Relation between soil erodibility based on wind tunnel tests and percentage of clods greater than 0.84 mm. in diameter in various soil classes.

Many factors, not all of which are associated with erodibility of soil by wind, influence the amount of erosion. Because of this, it seems best to express the erodibility in dimensionless form applicable to any set of conditions other than those of the soil itself. A convenient way of expressing erodibility on a dimensionless basis is by erodibility index $I$. This index is equal to $X_2/X_1$, in which $X_1$ is the quantity eroded when the soil contains 60 percent of clods greater than 0.84 mm. and $X_2$ is the quantity eroded under the same set of conditions from soil containing any other proportion of clods greater than 0.84 mm. in diameter. This is essentially the same as erodibility

index I used in previous publications (22, 23): $I = 10\dfrac{x(E)}{x(E)_{70}}$, in which $x(E)$ is the quantity of soil eroded and $x(E)_{70}$ is the quantity of soil eroded under the same set of conditions when the soil contains 70 percent of dry fractions less than 0.84 mm. in diameter. For any given soil the value of the erodibility index I will be about the same, irrespective of which wind tunnel is used in determining the erodibility. The relationship of erodibility index I to soil cloddiness as determined by dry sieving is shown in figure 17. The curve drawn through the average of individual measurement in figure 17 can be used to estimate approximately the erodibility index based on the percentage weight of clods greater than 0.84 mm. in the soil; do not consider, for the present, the influence due to differences in mechanical stability of the clods.

Some idea of the degree of error that is possible in erodibility index estimated from the percentage of clods greater than 0.84 mm. can be obtained by observing the magnitude of deviation of determined values of erodibility from the average curve shown in figure 17. Major sources of possible error in erodibility index estimated from the percentage of clods greater than 0.84 mm. are as follows:

(1) The order of erodibility on any group of related soils is usually the same, irrespective of the drag velocity to which they are subjected. On widely different soils the order might be reversed with a change of drag velocity. That is especially true when comparisons are made of extremely different soils, such as a fine sandy soil containing a preponderance of highly erodible fractions and a clay soil containing a large proportion of semierodible fractions.

(2) Erodibility is based on volume of nonerodible clods and not on weight, as determined for soils shown in figure 17. If the bulk density of the clods and the erodible fractions are the same, either the percentage volume or the percentage weight can be used with equal effect. However, if the two fractions have different densities, some error in the estimation may be expected.

(3) Differences in the size of clods have considerable influence on erodibility, but no distinction of size distribution of clods is made in figure 17. Clods 0.84 to 6.4 mm. in diameter, for example, are generally more than twice as effective in reducing erosion as clods 6.4 to 40 mm. in diameter (10).

(4) Erodibility is based on the equivalent diameter distribution of the erodible particles, not just on their proportion to the total weight of the soil. Determining the equivalent diameter distribution and estimating its influence on erodibility are quite laborious, however, and these calculations are probably not justified for the degree of refinement that will be obtained in the method of estimation.

A more exact, though more laborious, system of estimating erodibility of noncrusted cultivated soils is given in a separate publication (19).

A surface crust is invariably formed when the soil is wetted by rain and dried. The crust varies greatly in its resistance to erosion by wind, depending on the nature of the rain and the soil and the quantity of vegetative cover on the surface. Erodibility of a loose, freshly cultivated soil is usually reduced when the soil is wetted by rain and dried (table 6). In like manner, erodibility is generally increased when

the surface crust is destroyed, such as by abrasion from windborne materials. The surface crust is usually so weak it has virtually no influence on the size distribution of dry aggregates determined by dry sieving. The average ratio of erodibility of a crusted soil to erodibility of a noncrusted soil is about 1:6 (table 6). This ratio is in general agreement with that obtained on a large number of soils in a crusted and noncrusted condition reported in a previous publication (5).

If it is assumed that values of erodibility index I apply to loose, noncrusted soils as on freshly cultivated fields, the relative erodibility of soil whose surface is completely covered with a surface crust and has the same degree of cloddiness is about one-sixth I. Complete surface crusting usually occurs when a cultivated soil is first wetted and dried and before any erosion has taken place. However, there are all sorts of conditions of the surface crust between these two extremes, depending principally on soil texture and consequent erosional intensity since the last tillage operation. No manual or mechanical method has been devised on how to measure the degree of development of the surface crust. The only method available at the present time is based on a visual observation of the proportion of the original crust still remaining after weathering and erosion.

By taking cognizance of the usually variable status of the surface crust, the relative soil erodibility E at the time the estimation is made may be expressed by

$$E = (1 - bC)I \qquad (11)$$

where C is the percentage of the surface crust remaining after weathering and erosion and b is equal to 0.00833.

## Comparison of Estimated Erodibility With Natural Erodibility

Sixty-nine sites, representing as many fields, in western Kansas and eastern Colorado were chosen in 1954, 1955, and 1956 for the purpose of checking the validity of estimations of wind erodibility of soils in the spring, based on wind-tunnel tests. The quantity of natural erosion on each site was estimated visually, as shown in table 8.

The average erodibility computed from soil cloddiness, quantity of crop residue, and surface roughness in accordance with the previously described method (22) and the average quantities of natural erosion on three major groups of soil are shown in table 9. At the beginning of the spring season the order and the relative magnitude of computed and natural erodibility of the fields on different soil classes were about the same. Soil cloddiness, crop residues, and surface roughness changed little from the beginning to the end of the season. However, the natural amount of erosion increased greatly on fine sand and loamy fine sand, considerably on fine sandy loam, and only slightly on silt loam and silty clay loam soils as the season of high wind erosion came to an end. The sands were most susceptible to the abrasive action of windblown soil material. The surface crust and clods on this soil class were most fragile and disintegrated readily under abrasion. Next in order of resistance to abrasion were the loamy sands, then came the sandy loams, and then the loams, silt loams, and silty clay loams.

TABLE 8.—*Visual estimation of soil erodibility and of erodibility based on wind-tunnel tests from quantity of natural erosion in Kansas and Colorado, 1954–56*

| Quantity of erosion | Description of erosion | Erodibility based on wind tunnel tests (25) |
|---|---|---|
| | | *Tons per acre* |
| None | Insignificant; no visible effects of soil movement. | < 0.25 |
| Slight | Soil removal down to ¼ inch, not sufficient to kill wheat. | 0.25 to 1 |
| Moderate | Removal and associated accumulations ¼ to ½ inch deep, sufficient to kill wheat. | 1 to 5 |
| High | ½- to 1-inch removal and associated accumulations. | 5 to 25 |
| Very high | 1- to 2-inch removal and associated accumulations. | 25 to 125 |
| Exceedingly high | Greater than 2-inch removal with appreciable dune formation. | >125 |

The latter group of soils, which constitute most of the "hardlands," is probably the most resistant to the abrasive action of wind erosion. Their resistance is due to the ease with which silty clay loams are dispersed by water and their tendency to form a wind-resistant surface crust after they are wetted and dried. The relative amount of natural erosion increased over the computed amount inversely with the fineness of soil texture up to silty clay loam. Clays were not available for this study, but previous studies (15, 18) have indicated them to be about equal to fine sandy loam with respect to degree of cementation among the clods and abradability of the surface crust.

TABLE 9.—*Computed erodibility and quantities of natural erosion on 3 major groups of soil in Kansas and Colorado, 1954–56*

| Soil textural class | Computed erodibility, Mar. 15 | Average amount of erosion— | | Average condition of the surface crust— | |
|---|---|---|---|---|---|
| | | About Mar. 15 | About Apr. 30 | About Mar. 15 | About Apr. 30 |
| | *Tons per acre* | *Tons per acre* | *Tons per acre* | | |
| Fine sand and loamy fine sand. | 4.60 | 4.20 | 44.0 | 20 percent destroyed by erosion. | Almost all destroyed by erosion. |
| Fine sandy loam. | .65 | .95 | 2.0 | 10 percent destroyed by erosion. | 40 percent destroyed by erosion. |
| Silt loam and silty clay loam. | .19 | .14 | .19 | Intact. | Almost intact. |

These results showed that erosion once "broke loose," on sandy soils tended to destroy the surface crust and made the soil more erodible as the season progressed. Erosion on hardlands, on the other hand, was kept in check by limited quantities of loose soil material available on the surface of the ground and the limited effect of the loose material on the status of the surface crust. Toward the end of the season, therefore, the relative cumulated quantity of erosion varied from that obtainable on a fully crusted surface to that on a loose, noncrusted surface, depending on the mechanical stability of the surface crust and clods. If the soil had no surface crust, as in a freshly cultivated field, the quantities of natural erosion of the order of *I* applied. If, on the other hand, the soil surface was completely crusted, quantities of erosion were on the order of about one-sixth *I*, thereby confirming previous results on the relative influence of crusting as determined by wind-tunnel tests.

Results obtained with portable wind-tunnel tests in western Texas and other locations (22, 23) further confirmed the importance of soil surface crusting and mechanical stability of dry soil structure on erodibility by wind. The Texas tests were conducted on fields, some of which were highly eroded by preceding winds. The soil surfaces on fine sands and loamy fine sands were loose and noncrusted, those on fine sandy loams were generally partly crusted, and those on silt loams and silty clay loams were highly crusted. Soil erodibility based on wind-tunnel tests was therefore five-sixths *I* for fine sands and loamy fine sands, about one-half *I* for fine sandy loams, and about one-sixth *I* for silt loams and silty clay loams, other conditions remaining the same. In other tests where many of the sandy fields were not influenced by erosion and which therefore had a considerably developed surface crust (22), the erodibility index was one-half *I* for sandy soils and one-sixth *I* for the finer textured soils.

## Estimating Potential Erodibility

It is important to consider the magnitude of erosion that is likely to occur on soils of different textures and cloddiness if weather conditions become such as to make erosion possible. Erosion by wind has occurred in substantial parts of the southern Great Plains, 1952–56, inclusive. Under conditions of considerable erosion the fine sands and loamy fine sands had the surface crust and surface clods mainly destroyed and the quantities of erosion were of the order of *I* if crop residue and surface roughness remained the same. On silt loams and silty clay loams the surface crust and surface clods mainly were preserved, and the relative amounts of erosion were more on the level of one-sixth *I*. Other soils had the relative amounts of erosion somewhere between these two extremes.

The soil textural class serves as an index of resistance of clods and surface crust to disintegration by erosional abrasion, which is a contributing factor influencing the amount of natural erosion if and when it occurs. The potential erodibility can be determined from the generalized alinement chart of a previous publication (22), if the

erodibility obtained from that chart is multiplied by a factor depending on soil textural class as follows:

| Soil textural class: | Factor |
|---|---|
| Fine sand | 6 |
| Fine loamy sand | 4 |
| Fine sandy loam and clay (except saline clay) | 2 |
| Loam, silt loam, clay loam, or silty clay loam | 1 |

The erodibility values are indexes of erodibility of the soil surfaces and not the actual quantities erodible under field conditions.

# LITERATURE CITED

(1) BAGNOLD, R. A.
1943. THE PHYSICS OF BLOWN SAND AND DESERT DUNES. 265 pp., illus. William Morrow & Co., New York.

(2) BAVER, L. D.
1956. SOIL PHYSICS. Ed. 3. 489 pp., illus. John Wiley & Sons, Inc., New York.

(3) BRADFIELD, R., and JAMISON, V. C.
1938. SOIL STRUCTURE—ATTEMPTS AT ITS QUANTITATIVE CHARACTERIZATION. Soil Sci. Soc. Amer. Proc. 3: 70–76, illus.

(4) CHEPIL, W. S.
1941. RELATION OF WIND EROSION TO THE DRY AGGREGATE STRUCTURE OF A SOIL. Sci. Agr. 21: 488–507, illus.

(5) _____
1943. RELATION OF WIND EROSION TO THE WATER-STABLE AND DRY CLOD STRUCTURE OF SOIL. Soil Sci. 55: 275–287, illus.

(6) _____
1945. DYNAMICS OF WIND EROSION: II. INITIATION OF SOIL MOVEMENT. Soil Sci. 60: 397–411, illus.

(7) _____
1946. DYNAMICS OF WIND EROSION: IV. THE TRANSLOCATING AND ABRASIVE ACTION OF THE WIND. Soil Sci. 61: 167–177, illus.

(8) _____
1946. DYNAMICS OF WIND EROSION: V. CUMULATIVE INTENSITY OF SOIL DRIFTING ACROSS ERODING FIELDS. Soil Sci. 61: 257–263, illus.

(9) _____
1950. PROPERTIES OF SOIL WHICH INFLUENCE WIND EROSION: I. THE GOVERNING PRINCIPLE OF SURFACE ROUGHNESS. Soil Sci. 69: 149–162, illus.

(10) _____
1950. PROPERTIES OF SOIL WHICH INFLUENCE WIND EROSION: II. DRY AGGREGATE STRUCTURE AS AN INDEX OF ERODIBILITY. Soil Sci. 69: 403–414, illus.

(11) _____
1951. AN AIR ELUTRIATOR FOR DETERMINING THE DRY AGGREGATE SOIL STRUCTURE IN RELATION TO ERODIBILITY BY WIND. Soil Sci. 71: 197–207, illus.

(12) _____
1951. PROPERTIES OF SOIL WHICH INFLUENCE WIND EROSION: IV. STATE OF DRY AGGREGATE STRUCTURE. Soil Sci. 72: 387–401, illus.

(13) _____
1951. PROPERTIES OF SOIL WHICH INFLUENCE WIND EROSION: V. MECHANICAL STABILITY OF STRUCTURE. Soil Sci. 72: 465–478, illus.

(14) _____
1952. IMPROVED ROTARY SIEVE FOR MEASURING STATE AND STABILITY OF DRY SOIL STRUCTURE. Soil Sci. Soc. Amer. Proc. 16: 113–117, illus.

(15) _____
1953. FACTORS THAT INFLUENCE CLOD STRUCTURE AND ERODIBILITY OF SOIL BY WIND: I. SOIL TEXTURE. Soil Sci. 75: 473–483, illus.

(16) CHEPIL, W. S.
1953. FIELD STRUCTURE OF CULTIVATED SOILS WITH SPECIAL REFERENCE TO ERODIBILITY BY WIND. Soil Sci. Soc. Amer. Proc. 17: 185–190, illus.

(17) _____
1954. SEASONAL FLUCTUATIONS IN SOIL STRUCTURE AND ERODIBILITY OF SOIL BY WIND. Soil Sci. Soc. Amer. Proc. 18: 13–16, illus.

(18) _____
1955. FACTORS THAT INFLUENCE CLOD STRUCTURE AND ERODIBILITY OF SOIL BY WIND: IV. SAND, SILT, AND CLAY. Soil Sci. 80: 155–162, illus.

(19) _____
1956. INFLUENCE OF MOISTURE ON ERODIBILITY OF SOIL BY WIND. Soil Sci. Soc. Amer. Proc. 20: 288–292, illus.

(20) _____ and BISAL, F.
1943. A ROTARY SIEVE METHOD OF DETERMINING THE SIZE DISTRIBUTION OF SOIL CLODS. Soil Sci. 56: 95–100, illus.

(21) _____ and MILNE, R. A.
1941. WIND EROSION OF SOILS IN RELATION TO SIZE AND NATURE OF THE EXPOSED AREA. Soil Agr. 21: 479–487, illus.

(22) _____ and WOODRUFF, N. P.
1954. ESTIMATIONS OF WIND ERODIBILITY OF FIELD SURFACES. Soil and Water Conserv. Jour. 9: 257–265, illus.

(23) _____ WOODRUFF, N. P., and ZINGG, A. W.
1955. FIELD STUDY OF WIND EROSION IN WESTERN TEXAS. U. S. Dept. Agr. SCS–TP–125.

(24) CHILCOTT, E. F.
1937. PREVENTING SOIL BLOWING ON THE SOUTHERN GREAT PLAINS. U. S. Dept. Agr. Farmers' Bul. 1771, 28 pp., illus.

(25) COLE, R. C.
1939. SOIL MACROSTRUCTURE AS AFFECTED BY CULTURAL TREATMENTS. Hilgardia 12: 429–472, illus.

(26) GOLDSTEIN, S., editor.
1938. MODERN DEVELOPMENTS IN FLUID DYNAMICS. 702 pp., illus. (2 vol.) Clarendon Press, Oxford.

(27) HOPKINS, E. S., PALMER, A. E., and CHEPIL, W. S.
1946. SOIL DRIFTING CONTROL IN THE PRAIRIE PROVINCES. Canada Dept. Agr. Pub. 568 (Farmers' Bul. 32), 59 pp., illus. (Revised.)

(28) KEEN, B. A.
1933. EXPERIMENTAL METHODS FOR THE STUDY OF SOIL CULTIVATION. Empire Jour. Expt. Agr. 1: 97–102, illus.

(29) MANGELSDORFF, E. G.
1929. EXPERIMENTELLE BEITRÄGE ZUR BODENBEARBEITUNG. Landw. Jahrb. 69: 485–519, illus.

(30) MATHEWS, O. R.
1954. IMPLEMENTS AND METHODS OF TILLAGE TO CONTROL SOIL BLOWING ON THE NORTHERN GREAT PLAINS. (Revised.) U. S. Dept. Agr. Farmers' Bul. 1797, 21 pp., illus. (Original edition prepared by John S. Cole and George W. Morgan.)

(31) MECH, S. J.
1955. WIND EROSION CONTROL IN THE COLUMBIA BASIN. Wash. Agr. Expt. Stas. Sta. Cir. 268, 5 pp.

(32) NEKROSSOV, P. A.
1934. AGROPHYSICAL METHODS FOR EVALUATING THE WORK DONE BY CULTIVATION IMPLEMENTS. Internatl. Soc. Soil Sci. Trans., Sov. Sect. A, 2, 1: 44–56.

(33) OLMSTEAD, L. B.
1947. THE EFFECT OF LONG-TIME CROPPING SYSTEMS AND TILLAGE PRACTICES UPON SOIL AGGREGATION AT HAYS, KANSAS. Soil Sci. Soc. Amer. Proc. 11: 89–92.

(34) PUCHNER, H.
1911. AGGREGATE ANALYSIS ("SCHOLLENANALYSE") AS A MEANS FOR DETERMINING SOIL STRUCTURE. Deutsch. Landw. Gesell. Mitt. 26: 38–40.

(35) RICHARDS, L. A.
    1953. MODULUS OF RUPTURE AS AN INDEX OF CRUSTING OF SOIL. Soil
        Sci. Soc. Amer. Proc. 17: 321–323, illus.

(36) SIDDOWAY, F. H., MCKAY, H. C., and KLAGES, K. H.
    1956. DRYLAND TILLAGE METHODS AND IMPLEMENTS. Idaho Agr. Expt.
        Sta. Bul. 252, 46 pp., illus.

(37) SLATER, C. S.
    1951. WINTER ASPECTS OF SOIL STRUCTURE. Soil and Water Conserv.
        Jour. 6: 38–40, 42, illus.

(38) THROCKMORTON, R. I., and COMPTON, L. L.
    1937. SOIL BLOWING IN KANSAS AND METHODS OF CONTROL. Ia Soil
        Erosion by Wind in Kansas, Kans. Board of Agr. Rpt. 56, No.
        224–A, pp. 7–44, illus.

(39) VAN DOREN, C. E.
    1944. THE EFFECT OF CLODDINESS OF SOILS ON THEIR SUSCEPTIBILITY TO
        WIND EROSION. Amer. Soc. Agron. Jour. 36: 859–864.