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B. TRANSPORT BY WIND

CHAPTER 7

TRANSPORT OF SOIL AND SNOW BY WIND

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1. Introduction

Wind removes, deposits and mixes soil and snow. On soil, these actions are virtually permanent, but on snow they disappear with the snow.

In removal, the wind carries and scatters the topsoil hundreds of miles. In deposition, it forms the various types of aeolian materials representing extensive areas of loess soils in every continent. In mixing, it carries the soil across the land, creating surprising uniformity of minerals in the soil.

Movement of soil and snow by wind obeys essentially the same laws. Therefore, any field surface, crop strip, or wind barrier that reduces soil movement also traps snow. Although the methods to control soil and snow movement are basically the same, different ends are accomplished: they trap the snow that is initially in motion while they prevent the motion of soil particles that are initially at rest.

Wind acts on soil mostly in arid regions, but subsequent deposition extends, imperceptibly to a casual observer, even to humid regions (Malin, 1946; Waggoner and Bingham, 1961).

Although the wind is an agent of deposition and soil formation, it is also one of the great destroyers of soils. Generally, the wind acts on the soil like a fanning mill on grain—removing the finer and lighter constituents and leaving the coarser and denser ones behind. In many countries this sorting has transformed fertile soils to sandy wastelands. The downfall of ancient civilizations of central Asia, the Middle East, and North Africa is a record of depletion of grasslands and forests, wind erosion, and soil ruin (Jacks and Whyte, 1939; Bennett, 1939).

In the Great Plains of North America the most serious wind erosion occurred in the 1930's. The great economic losses caused by wind erosion stimulated serious attention to its causes, effects and remedies. Wind erosion control programs were established. Soil surveys and research on soil erosion were accelerated. Countless reports of research on wind erosion were published. This chapter is an attempt to summarize the research on the mechanics of soil and snow movement by wind and to indicate where progress still needs to be made.

The chapter deals with equilibrium forces between wind and soil and snow particles; with avalanching, sorting, abrasion, and deposition of transported particles and how these processes relate to principles of control; and finally with the wind erosion equation designed to evaluate methods of controlling wind erosion.

2. The surface wind

The structure of wind up to about 1.5 meters above the earth's surface affects the movement of soil and snow. A wind strong enough to move soil and snow is always turbulent with eddies moving at variable velocities and in all directions. The average forward velocity, generally regarded as velocity of a turbulent wind near the ground, increases with height according to a logarithmic equation of von Kármán (1934) and Prandtl (Brunt, 1944). Zero velocity exists somewhere above the average rough elements of the surface. The velocity above these projections and to at least 1.5 meters above the aerodynamic surface conforms with

$$u_1 = k^{-1}u_*\ln(z_1 z_0^{-1}). \tag{1}$$

in which u_1 is velocity at reference height z_1, u_* is the so-called friction velocity, k is the von Kármán constant with a value of 0.4, and z_0 is the roughness length.

The friction velocity, u_*, is an index of the rate of increase of velocity with height. The stronger the wind, the greater is u_*; but for a rigid surface, z_0 remains the same no matter how strong the wind. Moreover, u_* remains constant for a wind of a given speed as measured at some fixed height (Bagnold, 1943).

As soon as erosion sets in, velocity near the ground is reduced and a new drag velocity, u_*', becomes established (Bagnold, 1943):

$$u_1 = k^{-1}u_*'\ln(z_1 z_t^{-1}) + u_t \tag{2}$$

in which z_t is a height above the mean aerodynamic surface where all friction velocities, u_*', for different winds converge. u_t is velocity at reference height, z_t. u_t remains constant no matter how strong the wind and varies only with size and density of erodible soil or snow particles. The more erodible the surface, the greater is the concentration of jumping particles and the greater is the reduction of wind velocity near the ground.

For rough pipes and smooth soil surfaces, von Kármán (1934) and Nikuradse (1950) found that the mean wind frictional drag, $\bar{\tau}$, associated with a wind of a given speed measured at a fixed height remains constant no matter how high (up to 1.5 mm) the surface roughness. They found that

$$\bar{\tau} = \rho(u_*)^2 \tag{3}$$

in which ρ is the density of air. If u_* is expressed in cm sec^{-1}, then $\bar{\tau}$ is in dynes cm^{-2}. The drag $\bar{\tau}$ is dynamic, acting generally in the direction of flow.

However, for rough vegetation and bare soil surfaces with height of roughness varying from 5 to 50 cm, Sheppard (1947) and Chepil and Siddoway (unpublished data, 1962) found that $\bar{\tau}$ increases greatly with surface roughness. Therefore, Eq. (3) cannot be used to compute the surface drag from the friction velocity as generally has been done. Consequently, Eq. (3) must be modified to

$$\bar{\tau} = C_d\rho(u_*)^2 \quad \text{or} \quad C_d\rho(u_*')^2 \tag{4}$$

for noneroding and eroding surfaces, respectively, in which C_d is an empirical coefficient that varies with the nature and degree of surface roughness and which for a relatively smooth surface with roughness elements not exceeding 1.5 mm has a value of 1. Coefficients such as this are used in this chapter to represent the influence of little understood and unrecognized factors that influence the movement of solid particles by wind.

3. Mechanics of soil and snow transportation

A. Equilibrium Forces on Particles at Threshold of Their Movement

Movement of solid, loose, and dry particles begins when the pressures of the wind against the most erodible particles overcome their immersed weight. The wind exerts three pressures on a particle resting on the ground (Einstein, 1950; Ippen and Verma, 1953; Chepil, 1959). One is a positive pressure against that part of the particle facing into the wind; it is known as the impact or velocity pressure. The second is negative and on the lee side of the particle; it is known as viscosity pressure. The sum of the two pressures is the total drag, F_c, often referred to as drag. F_c acts in the general direction of the wind at about a level indicated in Fig. 1.

The third pressure which is negative and on top of the particle is caused by the Bernoulli effect. It is called static or internal pressure and causes a lift on the particle. Lift, L_c, generally acts at right angles to the direction of fluid motion. L_c acts through the center of gravity, c.g. (Fig. 1).

The immersed weight of a spherical particle is equal to $0.52gD^3\rho'$ in which g is gravity, D the grain diameter, and ρ' the immersed density of the particle (Fig. 1).

The threshold drag, F_c, and lift, L_c, required to move the topmost cohesionless particles lying on the surface are influenced by the diameter, shape, and immersed density of the particle; the angle or repose, Φ, of the grains with respect to the mean drag level of the wind (Fig. 1); the closeness of packing, η, of the top grains on the sediment bed; and the impulses of wind turbulence, T, associated with drag and lift (White, 1940; Einstein, 1950). From wind tunnel experiments on loose, dry soil particles, Chepil (1959) found that $L_c = 0.75\ F_c$, $\Phi = 26°$, $\eta = 0.2$, and $T = 2.5$, and by substituting, transposing, and factoring these values and those of Fig. 1 composed the threshold drag equation

$$\bar{\tau}_c = 0.66gD\rho' \tan\Phi\eta(1 + 0.75 \tan\Phi)^{-1}T^{-1} \tag{5}$$

in which $\bar{\tau}_c$ is the mean threshold drag per unit horizontal area of the ground. In the tunnel, the mean threshold drag, $\bar{\tau}_c$, for various types and sizes of soil particles greater than 0.1 mm in diameter agreed reasonably well with that computed from Eq. (5). This confirmed for loose, dry soils the general validity of Eq. (5) and its approximate parameters.

There is no apparent reason why (5) should not apply to loose, dry snow particles. Whereas ρ' in air for soil particles ranges from 1.5 to 2.7, for snow it is about 1.0; but Φ and η are probably not very different,

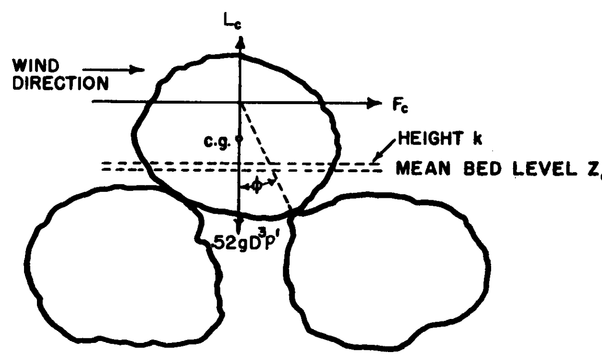

FIG. 1. Mean forces of drag, lift, and gravity acting on a particle greater than 0.1 mm in diameter in a windstream. Net moment opposing F_c is $(0.52gD^3\rho' - L_c)\tan\Phi$ for a spherical particle (Chepil, 1959, reprinted from *Soil. Sci. Soc. Amer. Proc.*, **23**, p. 424).

and T is expected to be about the same for both. Eq. (5) applies only to loose, dry, solid particles, whether they be soil or snow, and to particle diameters greater than about 0.1 mm. For snow, the threshold drag, $\bar{\tau}_c$, and the threshold drag velocity may well be lower than for soil owing to lower density of the particles (Seligman, 1936).

B. Equilibrium Forces During Particle Entrainment

The initial movement of soil and snow particles is in a series of jumps known as saltation (Fig. 2). The higher they jump, the more energy they derive from the wind. Particles subject to saltation generally range from 0.1 to 0.5 mm in equivalent diameter, $D_e = \rho' D(2.65)^{-1}$. If the wind is exceedingly strong, some particles as large as $D_e = 1$ mm may jump. Comparatively few soil particles jump higher than a meter. Over 90 per cent of them jump less than 30 cm (Chepil, 1945a). Snow crystals, on the other hand, have a lower density than soil and therefore jump higher and farther (Bagnold, 1943).

The forces of lift and drag on soil and snow particles change rapidly as the particles rise. Lift disappears a few centimeters above the ground. This height is considerably less than the height that many particles rise in saltation. Apparently inertia causes the particles to rise above the zone of lift. The greater the roughness and the friction velocity, the higher lift extends. On the other hand, drag on the particles increases with height as long as wind velocity increases with height (Chepil, 1961).

After being shot usually vertically or nearly so into the air, the particles stop because of gravity and fall at an accelerating velocity. At the same time drag accelerates them horizontally. The downward and forward accelerations are proportioned so that the inclined path of the falling particle is generally straight. However, drag is much greater than gravity and therefore the particle returns to the surface on a path 6 to 12 degrees from the horizontal. This angle depends on wind velocity and on equivalent diameter of the particle (Bagnold, 1943).

Movement in saltation causes two other movements—the rolling and sliding creep of coarser grains along the surface of the ground and the floating or suspension of fine dust particles.

The proportion of the three types of soil movement varies greatly for different soils. In the soils examined, 50 to 75 per cent of the weight of the soil was carried in saltation, 3 to 40 per cent in suspension, and 5 to 25 per cent in surface creep (Chepil, 1945a).

Although dust less than 0.01 mm in diameter is extremely resistant to movement by direct pressure

FIG. 2. Photograph of paths of particles moved from left primarily in saltation. The area photographed was 2 by 2 cm (Chepil and Woodruff, 1963, reprinted from *Advances in Agronomy*, **15**, p. 243).

of the wind, it is readily kicked up by larger particles moving in saltation, just as dust is kicked up by travelers. Once kicked into the air, dust particles can be lifted high in the atmosphere by eddies whose upward velocity of at least 100 to 150 cm per second is sufficient to lift to an indefinite height silt (0.002 to 0.02 mm), some very fine sand (0.02 to 0.1 mm), and snow crystals (up to about 0.2 mm). Dust clouds often rise 3000 to 4000 meters and are the most visible and therefore the most dramatic aspects of "dust storms." But dust clouds are only a show. The dominant process is the saltation of particles close to the ground (Chepil, 1958). Without saltation, dust clouds would never occur.

C. Rate of Soil and Snow Transportation

If the wind is greater than required to move the particles, then for dune sands (Bagnold, 1943) and dry soils (Chepil, 1958)

$$q = a\sqrt{D_e}(u_*')3 \tag{6}$$

in which q is weight of material that moves (per unit time) through a unit width of unlimited height normal to the direction of movement. It is logical to apply (6) to loose, dry snow. Eq. (6) shows that transportation varies directly as the cube of the drag velocity, u_*', and as the square root of the average equivalent diameter, D_e, of the transported particles.

The empirical coefficient a in Eq. (6) increases with the size of erodible particles (Bagnold, 1943), with the proportion and size of nonerodible fractions (Chepil, 1950), and with distance from the windward edge of the field (Agricultural Research Service, 1961). Coefficient a decreases with the proportion of fine dust particles (< 0.05 mm) (Chepil, 1958) and with moisture (Chepil, 1956). All these factors, and perhaps many more, affect the rate of soil and snow movement and hence coefficient a. Eq. (6) applies equally well to saltation, suspension, and surface creep, but coefficient a is different for each of these movements (Chepil, 1945b).

The rate of snow transportation is often lower than of soil because of greater cohesion caused by moisture films between the snow crystals (Bagnold, 1943). A surface crust forms more rapidly on snow than on soil (Seligman, 1936) and hinders or prevents further movement of both snow and soil.

Crusting of soils is caused primarily from wetting by rain or melting snow. On snow, crusting is primarily due to changes in temperature. During warm periods, a loose, coarsely crystalline condition of snow prevails. When frozen, these crystals form a crust from a fraction to several inches thick (Gerdel, 1945).

D. Dust Concentration, Visibility, and Rate of Removal

The weight of dust in unit volume of air varies with height as

$$C_z = bz^{-0.28} \tag{7}$$

in which C_z is concentration at height z above the ground and b is a coefficient which varies with the intensity of erosion (Chepil and Woodruff, 1957) Eq. (7) agrees reasonably well with the basic formula of Schmidt as reported by Vanoni (1946) in his experiments with particles suspended in water.

The dust suspended in the lower atmosphere can be estimated from daytime visibility and Fig. 3 (Langham *et al.*, 1938, and Chepil and Woodruff, 1957). The total dust load, C_m, refers to the cubic mile against the earth's surface and 1 mile high. For any intensity of erosion, the dust concentration and load above 1 mile can be estimated by Eq. (7).

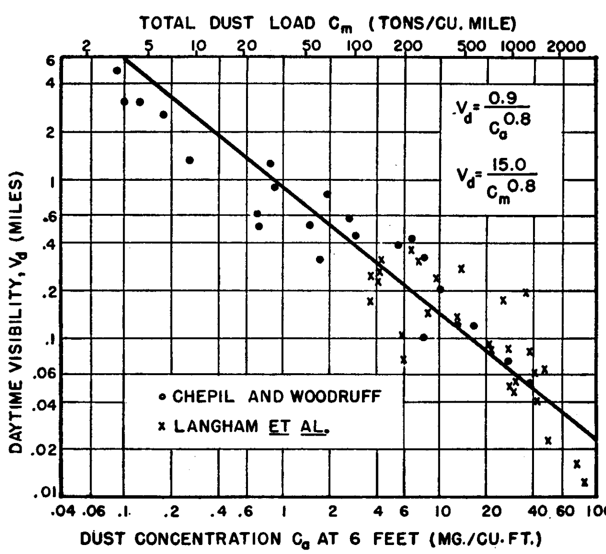

FIG. 3. Relation between daytime visibility and dust load and concentration (Chepil and Woodruff, 1957, reprinted from *Amer. J. Sci.*, **255**, p. 111).

Fig. 3 can also be used to estimate the amounts of dust removed from eroded regions. For this, it is necessary to know the number, dust concentration or visibility, duration, and wind velocity of dust storms. Through the vertical square mile nearest the earth and normal to movement, R in tons per hour (Chepil and Woodruff, 1957) is

$$R_m = 29.5\,V_d^{-1.25}(2.33m^{-1} + B) \tag{8}$$

in which V_d is daytime visibility of a dusty atmosphere in miles and m and B are coefficients which vary with wind velocity. A table of m and B for different wind velocities is given by Chepil and Woodruff (1957).

The rate of removal, R_m, times width of eroding region normal to wind direction times duration and number of dust storms occurring in a given time will give the quantity of dust removed from a region, assuming that the transport above 1 mile is negligible. In western Kansas and eastern Colorado during the 1954 and 1955 dust storms, the average rate of soil removal, R, was about 10,000 tons per hour per vertical square mile against the earth's surface (Chepil and Woodruff, 1957). The total duration of these dust storms was estimated from the weather records at Dodge City, Kansas, (which lies on the eastern outskirt of the severely wind-eroded area) to be 435 hours. Thus, during 1954 and 1955, 4.35 million tons of dust per mile width normal to wind direction moved past Dodge City and out of the eroded area. Assuming that this region is 400 miles wide along the direction of the wind and that an acre-foot of soil weighs 2000 tons, 0.1 inch of soil emigrated during 1954 and 1955. From the 1922-1961 weather records at Dodge City, the total duration of dust storms was estimated to be 5200 hours. Assuming that the intensity of dust storms was the same throughout the whole period, the net removal during the 40 years was 1.2 inches (3 cm) of soil. This estimate would be greater or smaller if we assumed the region to be narrower or wider than 400 miles along the wind direction, and greater if we took cognizance of dust carried above 1 mile.

Far more than the average of 3 cm of soil was removed from some eroded fields. For example, in one of the counties in Kansas, Chepil *et al.* (1952) found that during 20 years after breaking of virgin sod 30 representative fields had about 23 cm of topsoil removed. This is a tremendous rate of soil removal! Extensive areas of North America have been exposed to this removal, especially in the Great Plains.

E. Avalanching

Soil flow is zero on the windward edge of a level, eroding field but increases, or avalanches, for 10 to 500 meters or more before it becomes the maximum that a wind of a particular velocity can sustain (Agricultural Research Service, 1961). The maximum flow is approximately the same for all soils and is about equal to that of dune sand. Often the maximum flow is not reached because the distance across the eroding field is limited.

High soil erodibility, I, increases the rate of avalanching and decreases the distance in which maximum flow is reached (Table 1). Therefore, the more erodible the soil, the narrower the field or strip has to be to reduce erosion (soil flow) to some tolerable limit.

F. Sorting

The wind moves the finer and lighter particles faster than the coarser and denser ones (Chepil, 1957). The finer the eroded particles, the greater is their speed, height, and distance of travel.

The wind separates the soil into several distinct grades:

(a) *Residual soil materials*: Nonerodible clods and rocks.

(b) *Lag sands, lag gravels, and lag soil aggregates*: Semierodible grains that mainly creep over the surface and then rest here and there on eroded areas.

(c) *Sand and clay dunes*: Highly erodible grains moved in saltation and deposited far from an eroded area.

(d) *Loess*: Dust which was once lifted off the ground by impacts of saltating grains is carried aloft and deposited in uniform layers both near and far beyond the dunes. Dust is carried in true suspension. Freshly deposited dust resembles loess deposited in the Pleistocene (Swineford and Frye, 1945). The huge loess deposits in many parts of the world show the great importance of wind as a geologic force.

Table 1. Average distance in which maximum erosion (soil flow) on smooth, bare fields is reached.

Soil erodibility I* (tons/acre/annum)	Distance windward to reach maximum (meters)
220	250
134	650
86	1,200
56	2,000
38	3,000

* I is the average annual soil loss that would occur at Garden City, Kansas, on fields large enough for development of maximum soil flow.

There are no distinct demarcations among the various grades of wind-sorted materials. The size of one grade overlaps considerably with size of another.

In some cases wind erosion removes virtually all components of the surface soil (Chepil, 1957). This nonselective removal is associated primarily with loess which was already sorted during past geologic eras. Most soils, however, have not been deposited from the air, and here wind erosion removes slit, clay, and organic matter and makes the soil progressively sandier and less productive (Daniel, 1936).

G. Abrasion

Abrasion by the impact of particles blown along the surface occurs on all soils (Chepil, 1958). Soils are usually covered with a thin, resistant crust. As soon as some particles are loosened and moved by wind, their abrasion against the surface disintegrates the crust and exposes more erodible soil. Also, the nonerodible clods gradually break under impacts of saltating grains. The longer erosion continues, the greater is the quantity of erodible material abraded and removed. The materials abraded from clods and surface crust accumulate on the lee of fields or, if they are fine, travel far. The smaller the detached particles, the farther they are transported.

Movement of dry snow over bare or partly bare ground often contributes greatly to erosion of soil. Dry snowflakes usually change to particles large enough to move in saltation. The impacts of saltation break the nonerodible clods and surface crust to fragments which in turn are moved by wind. Thus, standing crop residues that trap the snow and prevent it from abrading the soil aid in controlling erosion (Hopkins *et al.*, 1946).

H. Deposition

Soil stabilization proceeds naturally or is accomplished by man in three successive stages: (1) deposition, (2) consolidation and aggregation of deposited erodible particles, and (3) revegetation of the surface.

The initial stage, deposition or stilling, can be accomplished in two ways. One is by sedimentation, i.e., the settling of soil particles through air that has slowed down. A familiar example is the settling of dust after a dust storm. The heavier the dust particles, the sooner and the closer to the eroded area they settle (Hanna and Bidwell, 1955).

The second way that erosion is stilled is by trapping, i.e., by stopping movement along the ground. Trapping may be accomplished by roughening the surface, by placing barriers in the path of the wind, by burying the erodible soil particles by tillage, or by revegetation. These methods will be discussed as means of control.

4. Principles of control

To control soil and snow movement, surface barriers and cover are employed (Chepil *et al.*, 1961). Perhaps the simplest way to explain this principle is to describe the behavior of bare soils subject to an erosive wind.

The removal of erodible particles continues until the height of the nonerodible clods is increased to a degree that completely shelters the erodible particles from the wind. Movement then ceases (Fig. 4).

When soil movement ceases, the ratio of downwind distance between the nonerodible barriers, D_s, divided by the height of the barriers, H, remains constant no matter what the ratio of erodible to nonerodible fractions in the soil. This is the critical surface barrier ratio. Although Chepil (1950) originally called this the critical roughness constant, the present term seems more appropriate. The ratio is in fact an effective distance between nonerodible surface barriers (measured in terms of heights of such barriers) required to reduce the quantity of erosion to zero.

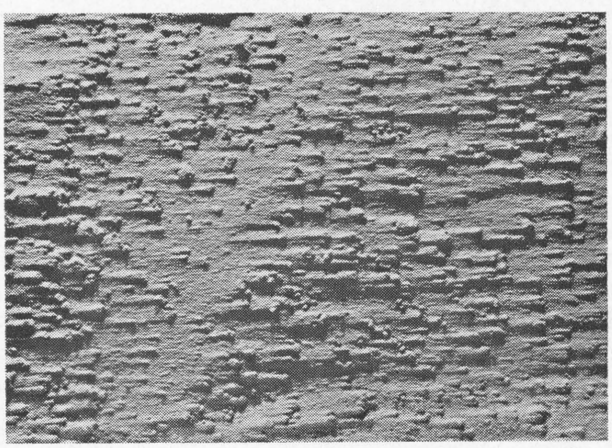

FIG. 4. Appearance of a silt loam composed of 92 per cent nonerodible fractions: a (upper), before exposure to wind, and b (lower), after exposure for the period required for soil removal to cease. Drag velocity of the wind was 60 cm per second and wind direction was left to right (Chepil, 1958, reprinted from *USDA Tech. Bull.* No. 1185, p. 12).

If the nonerodible barriers are extremely low, as in fine gravel, a relatively large number of the gravel pieces would be needed to protect the erodible particles from the wind for they would protect the erodible particles more by covering than by sheltering. Thus, virtually all nonerodible materials created on the surface of the soil or snow to control their movement have an element of cover in addition to barriers. The principle of surface barriers and that of cover is therefore inseparable; so also is the principle of soil consolidation and aggregation and that of barriers.

The principle of surface barriers and cover extends beyond the surface roughness elements composed of nonerodible clods. It extends to almost all elements employed in control, such as vegetative covers, nonerodible ridges, windbreaks, and crop strips. All devices are designed to:

(a) Take up force of the wind leaving little, if any, to be taken by the erodible particles.
(b) Trap any eroded snow or soil particles on the lee or among surface roughness elements or barriers, thereby reducing avalanching and intensity of erosion and preventing erosion from spreading.

The critical surface barrier ratio varied from 4 to 20 depending on the friction velocity of the wind, the threshold drag velocity of the soil, and the characteristics of the barriers (Chepil, 1950; Woodruff and Zingg, 1952).

A. VEGETATION AND OTHER TYPES OF COVER

The ultimate objective in any wind erosion control program is to cover the ground with vegetation or residue (Fig. 5). Vegetative cover, dead or alive, is

FIG. 5. This rodweeder with small duckfoot shovels leaves as much as 80 per cent of the wheat stubble mostly standing above the surface. It is an excellent implement for killing weeds and maintaining vegetative residues to control erosion and trap snow (Chepil and Woodruff, 1963, reprinted from *Advances in Agronomy*, **15**, p. 278).

one of the most effective, easiest, and most economical controls of soil and snow movement. Depending upon amount, type, and orientation of vegetative cover, 5 to 99 per cent of the wind force can be removed from the soil (Zingg, 1954). The finer, the more erect, and the greater the quantity of residue, the more wind force is taken up by the residue and the smaller the amount of erosion.

Twenty tons of fine, 50 of medium, and 100 of coarse gravel spread on an acre adequately controlled erosion of sandy soil where no traffic passed. Also, films of $\frac{1}{8}$ to $\frac{1}{4}$ gallon of resin, asphalt, and latex emulsions per square yard controlled erosion of dune sand, but at considerably greater expense than with vegetation either hauled in or grown in place (Agricultural Research Service, 1962).

B. Soil Clods and Ridges

Where insufficient vegetation exists, clods and ridges may be created by tillage to resist erosion temporarily. Tillage produces a cloddy surface by bringing to the surface compacted soil from below 2 to 4 inches. Clods that are just large enough not to be moved by wind (generally > 2 mm diameter) are most effective in protecting the erodible particles (Chepil, 1958).

If the soil is composed mostly of erodible particles, ridging does little because the ridges continue to erode. However, if the ridges contain a substantial proportion of nonerodible fractions, the erodible fractions move from the ridges, are trapped in the furrows, and the ridges are soon stabilized by a mantle of large clods (Woodruff *et al.*, 1957).

The effects of tillage are temporary because the forces of the weather tend to break the clods to sizes small enough to be moved by wind (Chepil, 1958). As the surface clods break, clods below the surface are formed, making repeated tillage necessary to maintain a cloddy surface.

C. Windbreaks

Windbreaks are 1 to 10 rows of trees and shrubs that provide a formidable barrier to the wind (Fig. 6). Other formidable barriers are crops in narrow rows, snow fences, solid wooden or stone fences, and earthen banks.

Nearly all barriers provide maximum reductions in wind velocity in their lee nearest to the barrier, with gradual decrease downwind. No matter how hard the wind blows, the percentage reductions for rigid barriers remain constant (Woodruff and Zingg, 1952); but when high wind blows through porous, resilient barriers, the percentage reductions increase slightly (Bates, 1944). Since the barrier functions as a snow trap and as a soil erosion control measure by reducing wind velocities below the threshold of initiating movement, the control is greater for low than for high winds (Theakston, 1962).

Protection extends furthest when wind blows perpendicular to the barrier length and is almost nil when the wind blows parallel. Dense barriers reduce velocity markedly over short leeward distances, whereas porous barriers provide smaller reductions but for more extended distances (Woodruff, 1954; Caborn, 1957). Generally, some porosity is desirable to gain extended protection; however, large openings cause air jetting and erosion. Narrow barriers are about as effective in reducing wind velocity as wider barriers of equal porosity (Bates, 1944; Nägeli, 1953).

In a wind tunnel, Woodruff and Zingg (1952) found that the critical barrier ratio for single- and multiple-row model windbreaks for an equivalent 40-mile-per-hour velocity was about 9. However, partial protection from wind barriers extends far beyond the critical barrier ratio. For example, Iizuka (1950) observed that a windbreak which apparently had a critical barrier ratio of about 9 for an unstated wind velocity decreased soil movement to 0.14, 18, and 50 per cent 10, 20, and 30 times the barrier height downwind. These rather limited influences of windbreaks emphasize the need for a system of windbreaks spaced at narrow intervals across the field (Fig. 6).

Unfortunately, windbreaks rob soil moisture from crops. In Colorado, Greb and Black (1961) found that roots extended to 2.5 times tree height substantially reduced the yield of wheat and sorghum to this distance. In Canada, where more snow was available, Staple and Lehane (1955) found that yield of wheat was reduced to about one tree height away from the trees. In the steppes of the USSR, yields were increased

FIG. 6. Windbreaks composed of several rows of trees or shrubs are effective in reducing wind velocities and controlling wind erosion for some distance to leeward from the windbreak (Chepil and Woodruff, 1963, reprinted from *Advances in Agronomy*, **15**, p. 285).

greatly between protective forest strips (Kalashnikov, 1952) apparently because snow was trapped and evaporation was reduced.

Plant barriers that compete as little as possible with crop growth are being sought. Tall annual crops interplanted in narrow rows between other smaller crops trapped drifting snow and prevented wind erosion (Sobolev, 1947). Rows of barley interplanted with asparagus protected the soil from wind (Schultz and Carlton, 1959). Rows of annual crops have also been used to trap snow and shelter transplanted tree seedlings (Ferber, 1958). The perennial grass, *Miscanthus*, planted in rows 15 to 30 meters apart controlled drifting sand along the coast of Taiwan (Sheng, 1961).

D. Crop Strips

Strips of erosion-resistant crops protected interplanted strips of erosion-susceptible crops. Protection of susceptible crops is not so much afforded by sheltering from the wind as by trapping of soil and snow and thereby reducing avalanching. The more erodible the soil, the narrower the erosion-susceptible strips have to be to keep erosion to some tolerable limit (Agricultural Research Service, 1961). The resistant strips must be wide enough to trap all the soil that may be moved in saltation. Unfortunately, stripcropping alone does not control erosion fully and must be supplemented by crop residues and soil clods for complete protection. However, less residue and clods will be necessary with than without stripcropping. Stripcropping adds assurance that wind erosion will not spread to adjacent fields, farms, and communities.

5. Wind erosion equation

An empirical equation with all its accompanying nomographs and tables has been developed to indicate approximate relations between wind erosion and the field and climatic factors that influence erosion (Agricultural Research Service, 1961; Chepil and Woodruff, 1963). It estimates (a) the potential wind erosion on any field under any climate, and (b) the surface roughness, soil cloddiness, vegetative cover, barriers, or width and orientation of field necessary to reduce the potential erosion to insignificance.

The equation embodies the primary factors that govern wind erodibility of land surfaces:

$$E = f(I,C,K,L,V) \qquad (9)$$

in which E, the potential average annual soil loss, is a function of: soil erodibility I, local wind erosion climatic factor C, soil surface roughness K, the maximum unsheltered distance across the field along the prevailing wind erosion direction L, and equivalent quantity of vegetative cover V. $L = D_f - D_b$ and $V = RSK_0$ in which D_f is total downwind distance across the field; D_b is the distance fully protected from wind erosion by a windbreak or barrier; R is the weight of vegetation above ground; S is a factor for kind of vegetative cover; and K_0 is a factor for orientation of the cover (whether standing, flattened, or leaning). Thus, eight specific data are needed to determine the potential soil loss of a field. The equation is mathematically complicated, but the accompanying nomographs and tables make estimations easy.

The equation can be worked in reverse to determine under any climate the condition of any factor required to reduce the potential loss, E, to any amount.

6. Needed research

More information is required on the influence of windbreaks and other barriers on air, soil, and snow movement; microclimate; and yields. Models and wind tunnels will speed this research.

Although we know what soil structure is ideal for resisting wind, we know little of how to create such a condition while permitting the soil to absorb water freely and serve as a good medium for plant growth. None of the present cropping and tillage systems are entirely suitable, and some are detrimental. New techniques for changing soil structure are needed.

Present methods of maintaining vegetation tend to leave the surface soil loose, fine, and highly erodible by wind. When drought occurs and vegetation becomes depleted, serious erosion may follow. Implements that create a resistant surface while maintaining vegetative residue on the surface need to be improved. We must learn how to develop vegetative matter resistant to decomposition and how to keep it above the ground. Plants for reclaiming eroding sand dunes are also needed.

Much damage to soils and crops could be avoided if severe wind erosion conditions could be predicted a few months or a year ahead of occurrence. Such predictions might be possible in view of severe wind erosion conditions tending to occur in cycles. A prediction would give opportunity to establish special tillage and cropping.

Many details lacking in the wind erosion equation should be filled as information is obtained. The wind erosion equation should be integrated with the water erosion equation, thus permitting complete predictions where we now have only fragmented information. Our success in this will reveal and summarize our understanding of erosion and how to control it most effectively.

7. Conclusion

Measurements showed that great quantities of dust have been removed by wind from dryland cultivated soils since they were broken from virgin sod. Records indicate, however, that the intensity of removal in some parts of the Great Plains during the 1950's was considerably less than during similar climatic conditions in the 1930's (Chepil and Woodruff, 1957; Chepil, Siddoway and Armburst, 1963; and unpublished data by Chepil *et al.*, 1963). This difference is believed to be due to better techniques of controlling wind erosion and to more earnest desire on the part of everyone to conserve the soil. Much greater strides, no doubt, are still to be made.

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