12

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Soil Conservation: Principles of Erosion by Wind

Interest in the fluid transport of solids—in this special case, wind erosion—continues not only because erosion affects soil productivity and damages plants but also because wind erosion is an important contributor of atmospheric aerosols.

The most comprehensive summaries on the movement of surface material by wind action have been done by Bagnold (1941) for desert sands and Chepil and Woodruff (1963) for agricultural lands. Control strategies have been discussed by Woodruff et al. (1977).

12-1 MEAN VELOCITY PROFILES

The fluid flows of interest to the agricultural community—those sufficient to cause wind erosion—are classified as turbulent boundary layer flows over relatively rough surfaces. For such flows over stable (noneroding) surfaces in the "constant stress layer" (the lower 10 to 20% of the boundary layer depth), the following form of the logarithmic law is often used to describe the mean velocity profile:

$$\bar{u}_z = \frac{u_*}{k}\left[\ln\left(\frac{Z - D}{Z_0}\right) + \phi_{(z)}\right] \qquad [1]$$

where $\bar{u}_z$ is the mean velocity at height z from some reference plane; u_* is the friction velocity defined as $(\tau_0/\varrho)^{1/2}$, where τ_0 is the shear stress at the surface and ϱ is air density; k is von Karman's constant (0.4); D is a zero

plane displacement; Z_0 is a roughness parameter; and $\phi_{(z)}$ is the integral diabatic influence function (Stearns, 1970). For most wind tunnel flows and winds of high velocity and turbulence, we can safely assume that near the surface $\phi_{(z)} = 0$ (the function $\phi_{(z)}$ is zero for adiabatic conditions).

For flows over relatively smooth—yet aerodynamically rough—stable surfaces, D is omitted because it becomes very small relative to Z and $Z_0 = 1/30\ d_p$ (Nikuradse, 1950), where d_p is particle diameter, so that Eq. [1] becomes

$$\frac{\bar{u}_z}{u_*} = \frac{1}{k} \ln \left(\frac{Z}{d_p} \right) + 8.5, \qquad [2]$$

which is also found in the literature.

Particles in transport near the surface alter the mean velocity profile. Based on work by Bagnold (1941) and Andres (1970), White and Schulz (1977) gave the following form of the profile equation for particle-laden flows:

$$\frac{\bar{u}_z}{u_*} = \frac{1}{k} \ln \left(\frac{Z}{d_p} \right) - 2.29 + 10.79 \frac{u_{*t}}{u_*} \qquad [3]$$

where u_{*t} is the threshold friction velocity. If u_* approaches u_{*t} (the stable surface case), Eq. [3] becomes Eq. [2].

12–2 PARTICLE DYNAMICS

Most published information concerning the movement of particles by wind distinguishes between three principal types of particle movement: suspension, saltation, and surface creep. However, the way the first particles are moved has received less attention. Before 1962, most writers were satisfied by Bagnold's (1941) description of particles rolling along the surface by direct wind pressure for about 30 cm before starting to bounce off the ground. Bisal and Nielsen (1962) concluded, after observing particles in a shallow pan mounted on the viewing stage of a binocular microscope, that most erodible particles vibrated (oscillated) with increasing intensity as wind speed increased and then left the surface instantly (as if ejected). Lyles and Krauss (1971), from wind tunnel observations, reported that as mean wind speed approached the threshold value, some particles (0.59 to 0.84 mm in diameter) began to rock back and forth (oscillate) and hypothesized that the oscillation frequency was related to the frequency band containing the maximum energy of the turbulent motion. Although average peak frequency of the longitudinal energy spectra and particle oscillation frequency were of same order of magnitude, more comprehensive research on particle oscillation is needed. Interestingly, Azizov (1977), who was studying soil moisture effects on wind erosion, noted a critical wind speed at which individual particles began oscillatory forward motions to a distance less than or equal to their diameters.

12-2.1 Suspension

Suspension of particulates by wind erosion often causes loss of productivity on the eroding field (Lyles, 1975), transport causes visibility and health hazards, and deposition causes chemical and sediment pollution. The suspended particles can range in size from about 2 to 100 μm, with mass median diameter of about 50 μm in an eroding field (Chepil, 1957a; Gillette and Walker, 1977). However, in long-distance transport, particles less than 20 μm in diameter predominate, because the larger particles have significant sedimentation velocities (Gillette, 1977). Some suspension-size particles are present in the soil, but many are created by abrasive breakdown during erosion. Some of the smallest are carried in crevices on the larger particles and are shed on impact (Rosinski et al., 1976).

Vertical flux (F_v) of suspension-size particles is often described by

$$F_v = F_0(u_*/u_{*0})^P \qquad [4]$$

where u_* is friction velocity, F_0 is a reference flux at a reference friction velocity u_{*0}, and P is a power that can range from about 2 to 6 (Englemann and Sehmel, 1976). Some evidence shows that P is about 3 on sandy soils (i.e., vertical flux increases at the same rate as horizontal), but P may be larger on finer-textured soils (Gillette, 1977). The value of F_0 is also uncertain. Chepil (1945) reported that 3 to 38% of the eroding soil could be carried in suspension, depending on soil texture. Generally, the vertical flux is less than 10% of the horizontal (Gillette, 1977, 1978).

Concentration profiles of suspended particulates can be easily predicted only for equilibrium conditions. When net vertical flux is zero,

$$F_v = -V_s\chi(Z) - K(Z)\,\partial\chi/\partial Z = 0 \qquad [5]$$

where V_s is settling velocity and χ is concentration. If particulate eddy diffusivity (K) is equal to that for momentum, $K = u_*kZ$, where $k \cong 0.4$ and Z is height above the surface, the solution of Eq. [5] is

$$\chi = \chi_1 (Z/Z_1)^{-V_s/u_*k}. \qquad [6]$$

Measurements during soil erosion confirm that these concentration profiles frequently occur (Chepil and Woodruff, 1957).

Widespread wind erosion causes dust storms, whose climatology in the contiguous USA has been studied by Orgill and Sehmel (1976). Highest dust frequency (with visibility less than 11 km) occurs in the Southern Great Plains, where most of the area is affected by dust 1% of the time. A small part of the area is affected by dust 3% of the time. Secondary dust frequency maxima occur in the western states, Northern Great Plains, Southern Coastal Pacific and inland valleys, and the Southeast. Highest frequencies occur in the afternoon between 1200 and 2000 local standard time. In the Great Plains, Hagen and Woodruff (1973) found the average dust

storm to last 6.6 h and estimated the median dust concentration to be 4.83 mg/m³ at observer height. Dusty hours vary widely from year to year, but the 2-year mean precipitation was correlated ($r^2 = 0.76$) with the annual hours of dust in the Great Plains. Pollard (1977), who investigated the meteorology of Southern Great Plains dust storms, found that wind speed and direction were correlated with their occurrence but that antecedent moisture was not.

From visibility and wind speed data, approximate calculations showed that in the Great Plains during the 1950's, 221 million t of dust was suspended annually and in the 1960's, 70 million t (Hagen and Woodruff, 1975). These calculations correlated well with dust depositions measured by Smith et al. (1970). These results suggest that particulate suspension from wind erosion exceeds that from all other sources (both natural and artificial) —the latter source accounts for about 31.9 million t annually in the USA (Walther, 1972).

The primary ambient air quality standard for total suspended particulates is 75 μg/m² (annual geometric mean) and 260 μg/m³ (maximum 24-hour average) (*Federal Register,* 1971). Because of the high concentration of coarse particles in dust storms, rural areas often fail to meet this standard (USEPA, 1976), and several investigators (e.g., Corn, 1971; Husar, 1976) have suggested that they be changed to reflect the health hazards posed by the size and chemical composition of the particulates.

Suspended particulates are best reduced by good wind erosion control practices. Tall vegetation can sometimes be used to trap particulates already suspended (Hagen and Skidmore, 1977; Smith, 1977).

12–2.2 Saltation

The characteristics of saltation (jumping) particles in wind have been described by several research workers (Free, 1911; Bagnold, 1941; Chepil, 1945; Zingg, 1953). However, theoretical understanding of the saltation process is at present incomplete (White and Schulz, 1977). In saltation, individual particles lift off the surface (eject) and follow distinctive trajectories under the influence of air resistance and gravity. Such particles (100 to 500 μm) rise at fairly steep angles but are too large to be suspended by the flow; they return to the surface either to rebound or to embed themselves and initiate movement of other particles. The bulk of total transport, roughly 50 to 80%, is by saltation. Saltating particles rise less than 120 cm; most rise less than 30 cm.

Chepil (1945) reported lift-off angles of 75 to 90° and impact angles of 6 to 12° from the horizontal. White and Schulz (1977), using microbeads to study particle trajectories in a wind tunnel, reported an average lift-off angle of 50 ± 20° for 57 filmed observations and average impact angle of 14 ± 3° for 43 observations.

Many investigations have shown the importance of lifting forces in saltation (White and Schulz, 1977). Because of the large differences between densities of air and sand (soil) particles, Owen (1964) suggested that the lift force due to interaction between particle motion and environmental vortici-

ty does not play a significant part in determining the motion of a particle (see Tennekes and Lumley, 1972, for the role of vorticity in turbulence dynamics). He did note that a lift force could result from the spin of a particle. White and Schulz (1977) have shown that considering the lift associated with particle rotation—called the Magnus effect—greatly improves the agreement between theoretical trajectories calculated from the equations of particle motion and those filmed in a wind tunnel. From photographs of saltation flow, Chepil (1954) found appreciable particle rotation of 200 to 1000 revolutions/s. From changes in light reflection of microbeads on film, White and Schulz (1977) estimated spinning rates of 115 to 500 revolutions/s.

Chepil (1945) reported that ratios of height of rise (h) to length of path (L) for saltating soil particles were about 1:10 for all the agricultural soils studied. From solutions of the equations of particle motion, White and Schulz (1977) indicated that h/L is a function of particle diameter (at constant u_*), with larger ratios associated with the smaller diameters in the saltation size range and decreasing to approach those of Chepil at 500 μm.

Finally, Owen (1964) stated that Bagnold deduced that lift-off velocities ($\bar{u}_L$) of saltating particles were comparable with u_*. White and Schulz (1977) reported $\bar{u}_L = 69.3$ cm/s when $u_* = 39.6$ cm/s, although about 18% of the particles examined from filmed trajectories had velocities between 20 and 40 cm/s. Thus, Bagnold's deduction is not unrealistic based on limited data.

12–2.3 Surface Creep

Mineral soil (sand) particles 500 to 1000 μm in diameter, too large to leave the surface in ordinary erosive winds, are set in motion by the impacts of saltating particles. In high winds, the whole sand surface appears to be creeping slowly forward at speeds $\ll 2.5$ cm/s—pushed and rolled (driven) by the saltation flow. The rippling of windblown sand has been attributed to unevenness in surface creep flow (Bagnold, 1941). Reportedly, surface creep constitutes 7 to 25% of total transport (Bagnold, 1941; Chepil, 1945; Horikawa and Shen, 1960). Factors affecting the proportion of total transport in surface creep seem to be wind speed, particle size distribution, and surface geometry (roughness).

12–3 THRESHOLD CONDITIONS

Bagnold (1941) used an experimental coefficient, A, to describe the threshold friction velocity, u_{*t} (the minimum at which the flow has sufficient energy to initiate particle movement). The equation is

$$u_{*t} = A (\alpha g\, d_p)^{1/2} \qquad [7]$$

where α is immersed density ratio, $(\varrho_p - \varrho)/\varrho$, g is the gravitational acceleration, d_p is particle diameter, ϱ_p is particle density, and ϱ is air density. In air, A has a value of 0.08 to 0.12, perhaps as great as 0.2 without salta-

tion flow. Iversen et al. (1973) and Wood et al. (1974) noted that A is a function of the particle friction Reynolds number $R_f = u_* d_p/\nu$, where ν is kinematic viscosity of air. In Wood's summary of previous research, A ranged from 0.08 to 0.22 for R_f greater than 0.7; however, most values were between 0.08 and 0.12. Iversen et al. (1976) extended Bagnold's equation to include interparticle forces (due to moisture, electrostatic effects, and other forces of cohesion) for estimating threshold values for small particles and in low-density atmospheres (extraterrestrial). Their estimates resulted in somewhat lower minimum values for u_{*t} in the fine particle range than those given by Eq. [7]. Those estimates and recent low air-density experiments (unpublished) also showed that the coefficient A is not a unique function of R_f for small particles, but because of interparticle forces, it is smaller for larger values of α. Some question remains concerning what value to use for d_p in materials of wide particle size range. Also, no standard method of determining u_{*t} has been used or specified. Theoretically, $u_{*t} \equiv u_{*\max}$ for $q \equiv 0$ (where q is particle flux), but it is difficult to measure experimentally (Lyles, 1977).

We suggest that threshold consideration be limited to flow and particle properties and exclude boundary geometry. There is a growing tendency to extend the threshold concept to include the presence of nonerodible roughness (elements)—thus making it dependent on many physical dimensions and geometric patterns of those elements. A clearer approach would be to partition the shear stress between erodible and nonerodible material and thus avoid confounding the threshold concept.

12-4 PARTICLE FLUX

When considering particle flux (flow rates), we must distinguish between cases in which all the erodible particles are less than 1000 μm and cases in which there are mixtures of erodible particles and nonerodible elements, e.g., aggregates or clods greater than 1000 μm and vegetative materials. Before we proceed, however, a brief review of soil avalanching (Chepil, 1957b) is appropriate.

On an unprotected eroding field, the particle flux is zero on the windward edge and increases with distance downwind until, if the field is large enough, the flux reaches the maximum amount that a given wind can sustain. This increase of particle flux with distance downwind has been called "soil avalanching." The more erodible the surface, the shorter the distance in which maximum flux is reached.

12-4.1 All Erodible Particles

The simplest case involves all erodible particles where several equations have been developed to predict soil (sand) flux from an area under specific soil and wind conditions. Most equations, empirically developed, relate mass of soil moved to surface-shear stress or friction velocity of the wind and erodibility characteristics of the soil.

The functional form of those equations is

$$q = f[(\text{soil properties})^a, (\text{flow properties})^b] \qquad [8]$$

where q is particle flux (mass per unit width per unit time) and a and b are constants. The soil properties term may include size, density, and shape; the flow properties term may include mean wind speed at some reference height ($\bar{u}_z$) or more commonly the friction velocity (u_*) for the flow in question, air density, and, rarely, a turbulence parameter. Specific flux equations for all erodible particles are contained in reports by O'Brien and Rindlaub (1936), Bagnold (1941); Kawamura (1951), Zingg (1953), Owen (1964), Kadib (1965), Makaveev (1967), and Iversen et al. (1976). The flux varies considerably among the equations because of different values found for the constants and for coefficients introduced in explicit equations. Because it contains a turbulence term, the Makaveev's (1967) equation is the only one given here:

$$q = C \, \varrho_p \, d_p \, \frac{\sigma_u}{\bar{u}_t} \left(\frac{\bar{u}_z}{\bar{u}_t} \right)^2 (\bar{u}_z - \bar{u}_t) \qquad [9]$$

where $\bar{u}_t$ is threshold mean wind speed at reference height z, σ_u is the root mean square (RMS) of the longitudinal velocity fluctuations (turbulence), and other terms are as previously defined. Substituting Eq. [1] into Eq. [9] and reducing give

$$q = C \, \varrho_p \, d_p \, \frac{\sigma_u}{u_{*t}} \left(\frac{u_*}{u_{*t}} \right)^2 (u_* - u_{*t}), \qquad [10]$$

which shows the well-known cubic relationship between friction velocity and particle flux and further supports why u_* is so often used to indicate the wind's capability to erode soil particles. Substituting the relationship $\sigma_u = C_1 u_*$ (Lumley and Panofsky, 1964) into Eq. [10] gives

$$q = C_2 \, \varrho_p \, d_p \left(\frac{u_*}{u_{*t}} \right)^3 (u_* - u_{*t}), \qquad [11]$$

which indicates particle flux is proportional to u_*^4 and illustrates the ability of u_* to characterize some turbulent properties of flow. All the previous particle flux equations assume steady or stationary *mean* flow. Fan and Disrud (1977) suggested that mean wind velocities are not generally stationary and that steady-flow flux equations may underpredict actual erosion.

12–4.2 Mixtures of Erodible Particles and Nonerodible Elements

Field soils seldom contain only erodible-size particles; mixtures of erodible and nonerodible elements are the more common and more complex case. An equation for transport of field soils is complicated by factors other

than erodible particle size gradation, e.g., proportion and size of nonerodible fractions, field roughness, vegetation, and soil moisture content. No reliable equations have been developed to express field soil flux for short time periods or for single erosive windstorms.

The protective role of nonerodible elements in the erosion process has been characterized by Lyles et al. (1974) in wind tunnel studies. For example, a soil initially with buried nonerodible elements like clods is eroded by a wind of characteristic friction velocity, u_*. As the erodible material is removed, the roughness is increased (because of exposure of portions of the nonerodible material), which increases the total friction velocity. The friction velocity (or drag) may be thought of as being divided between the nonerodible elements and the erodible soil (intervening surface). Thus, as more erodible material is removed from the initially smooth, erodible surface, more drag is absorbed by nonerodible elements and less is absorbed by the erodible soil. After sufficient time, enough soil is eroded so that the intervening surface drag decreases to the threshold where erosion ceases and the soil is stabilized. Stabilizing agricultural fields by nonerodible elements is complicated by variation in speed, direction, and duration of winds plus possible generation of erodible-size particles from larger aggregates by abrasion. However, the role of nonerodible elements is clearly to absorb part of the total wind drag—reducing the drag on erodible particles.

From wind tunnel studies, Lyles and Allison (1976) published a regression equation that predicts the degree of protection provided by standing crop residues and nonerodible soil aggregates:

$$\left(\frac{u_*}{u_{*t}}\right)_s = 1.638 + 17.044\,\frac{NA_s}{A_t} - 0.117\,\frac{L_y}{L_x} + [(1.0236)^c - 1] \qquad [12]$$

where $(u_*/u_{*t})_s$ is called the critical friction velocity ratio, because erosion begins when this value is exceeded; the larger the ratio, the greater the wind erosion protection. The term u_* is the total friction velocity for a stable surface at a given free stream velocity, and u_{*t} is the threshold friction velocity for the erodible particles in question. The other parameters are N/A_t, number of stalks in area A_t; A_s, average silhouette area (projected area facing flow) of a single stalk; L_y, distance (center to center) between stalks normal to wind direction; L_x, the corresponding distance in the wind direction; and c, percentage of dry soil aggregates greater than 1.0 mm in diameter (limited to $\leq 50\%$). Research is needed to test Eq. [12] under field conditions.

12-5 PRINCIPLES AND GENERAL STRATEGIES OF CONTROL

Two general principles of wind erosion control are obvious: (i) reduce wind forces on erodible particles or (ii) create particles resistant to wind forces. From knowledge of erosion processes and mechanics, four specific principles of wind erosion control have been identified (Woodruff et al., 1977):

1. Establish and maintain vegetation or vegetative residues.
2. Produce or bring to the soil surface nonerodible aggregates or clods.
3. Reduce field width along prevailing wind erosion direction.
4. Roughen the land surface.

These principles are unchanging, but the control *practices* that grow out of them (which are discussed in subsequent sections) vary in space (from place to place) and may change over time along with cropping and management systems. Any present or proposed (future) control strategy should be evaluated on how it influences the four principles. Those four principles plus a factor for climate were used by Woodruff and Siddoway (1965) to develop the now well-known wind erosion equation that predicts potential annual erosion rates:

$$E = f(I, K, C, L, V) \qquad [13]$$

where E is the potential annual soil loss rate, I is the soil erodibility, K is the soil ridge roughness factor, C is the climatic factor, L is the unsheltered median travel distance of the wind across a field, and V is the equivalent vegetative cover. Briefly, the soil erodibility index I is the potential soil loss in $t(ha\ y)^{-1}$ from a wide unsheltered, isolated field with a bare, smooth, non-crusted surface based on climatic conditions near Garden City, Kansas. It is related to surface soil cloddiness greater than 0.84 mm and may be de-termined by dry sieving and use of a table given by Woodruff and Siddoway (1965). Values range from 0 to 695 t/ha as surface aggregates greater than 0.84 mm decrease to 1%. For convenience, the U.S. Soil Conservation Service uses wind erodibility groups based primarily on soil texture as a guide for selecting I factors.

Soil ridge roughness factor K is a measure of the effects of ridges on erosion amounts relative to a smooth surface. It is determined from the height-spacing measurement of ridges caused by tillage implements and then consulting a chart given by Woodruff and Siddoway (1965). It varies from 0.5 for ridged surfaces to 1.0 for smooth surfaces.

The local wind erosion climatic factor C characterizes the erosive po-tential of climate (wind speed, precipitation, and air temperature) at a particular location relative to Garden City, Kansas, which has an annual value of 100% based on long-term data. Equations for computing the factor and a map giving general ranges of annual values for the western half of the USA are given by Chepil et al. (1962). Monthly climatic factors for most of the USA are presented by Skidmore and Woodruff (1968).

The equivalent field width L in the equation recognizes that the rate of soil flow increases with distance downwind across an eroding field until reaching the transport capacity of a given wind and that winds have a pre-vailing direction and a preponderance in the prevailing direction. It may be determined from field size and orientation and data on prevailing wind direction and preponderance in the prevailing wind erosion direction given by Skidmore and Woodruff (1968).

The vegetative cover factor V considers the quantity, kind, and orienta-tion of vegetation or vegetative residues on erosion amounts. All vegetative

materials must be converted to an equivalent small-grain standard before use in the equation. The standard is absolutely flat, small-grain stubble with straw aligned parallel with wind direction on smooth ground and in rows 25 cm apart at right angles to wind direction. Data for converting various crops to their small-grain equivalent are given by Craig and Turelle (1964), Woodruff and Siddoway (1965), and Lyles and Allison (1980, 1981).

Relations among variables in the equation are complex, and a single equation that expresses E as a function of the dependent variables has not been devised. The equation was initially solved in a stepwise procedure involving tables and graphs (Woodruff and Siddoway, 1965; Skidmore and Woodruff, 1968), but its solution has been computerized (Skidmore et al., 1970).

The equation was developed to estimate the potential erosion from a particular field or to estimate the field conditions necessary to reduce potential erosion to tolerable amounts and has been used extensively for design and evaluation of control systems (practices).

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