

4. Basic Wind Erosion Processes

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ABSTRACT

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Basic wind erosion processes are discussed under the major headings of soil particle dynamics, particle flow rates, and principles and general strategies of control. Particle dynamics are described in terms of suspension, saltation, surface creep, abrasion, sorting and threshold conditions. Soil particle flow rates are divided between the all-erodible-particle case and the more common but more complex case of mixtures of erodible particles and non-erodible elements. Specific principles of wind erosion control are identified. A wind erosion equation, which estimates potential erosion from a particular field, and the conditions necessary to reduce potential erosion to tolerable amounts are discussed.

INTRODUCTION

Wind erosion damages soil, crops and the environment by reducing soil productivity, affecting plant emergence, quality and yield, and increasing atmospheric particulates.

Conditions conducive to wind erosion exist when (1) the soil is loose, dry and finely granulated, (2) the soil surface is smooth and vegetative cover is sparse or absent and (3) the susceptible area is sufficiently large. These conditions often prevail in semiarid and arid climates, e.g. west of the 99th meridian in the U.S.A.

The most comprehensive summaries on the movement of surface material by wind have been done by Bagnold (1941) for desert sands and Chepil and Woodruff (1963) for agricultural lands. A review of the distribution of wind erosion on a global scale has recently been completed by Zachar (1982).

MEAN WINDS

Wind, of course, is the driving force behind all soil blowing. Winds sufficient to cause wind erosion are classified as turbulent boundary layer flows over

relatively rough surfaces. For such flows over stable (non-erodible) surfaces in the 'constant stress layer' (the lower 10–20% of the boundary layer depth), the following form of the logarithmic law is often used to describe the mean velocity profile:

$$\bar{U}_z = \frac{U_*}{k} \left[\ln \left(\frac{Z-D}{Z_0} \right) + \Phi_{(z)} \right] \quad (1)$$

where \bar{U}_z is the mean windspeed at height Z from some reference plane, U_* is the friction velocity defined as $(\tau_0/\rho)^{1/2}$, 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 eqn. (1) becomes

$$\frac{\bar{U}_z}{U_*} = \frac{1}{k} \ln \left(\frac{Z}{d_p} \right) + 8.5 \quad (2)$$

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:

$$\frac{\bar{U}_z}{U_*} = \frac{1}{k} \ln \left(\frac{Z}{d_p} \right) - 2.29 + 10.79 \frac{U_{*t}}{U_*} \quad (3)$$

where U_{*t} is the threshold friction velocity. If U_* decreases to U_{*t} (the stable surface case), eqn. (3) becomes eqn. (2).

PARTICLE DYNAMICS

Wind erosion consists of initiation, transport (suspension, saltation and surface creep), abrasion, sorting and deposition of soil aggregates and primary particles. The way the first particles are moved has received less attention than the modes of transport. 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 (in saltation). 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 windspeed increased and then left the surface instantly (as if ejected). Lyles and Krauss

(1971), from wind tunnel observations, reported that as mean windspeed approached the threshold value, some particles (0.59–0.84 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 the same order of magnitude, more comprehensive research on particle oscillation has not been reported. Azizov (1977), who was studying soil water effects on wind erosion, noted a critical windspeed at which individual particles began oscillatory forward motions to a distance less than or equal to their diameter.

Suspension

Suspension refers to the vertical and (eventual) horizontal transport of very small soil particles that are generally removed from the local source area. They may end up on the neighbor's farm or several states downwind. These particles are the showy part of wind erosion, seen by the observer as dust storms, and collectively have been called black blizzards, bread-basket dust and 'Kansas Grit'. 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 $< 20 \mu\text{m}$ in diameter predominate because the larger particles have significant sedimentation velocities (Gillette, 1977). Some suspension-size particles are present in the soil, but most are created by abrasive breakdown during erosion. Because organic matter and some plant nutrients are usually associated with the finer soil fractions, suspension samples are enriched in such constituents compared with the bulk soil source. Chepil (1945) reported that 3–38% of the eroding soil could be carried in suspension, depending on soil texture. Generally, the vertical transport is $< 10\%$ of the horizontal (Gillette, 1977, 1978).

Widespread wind erosion causes dust storms, whose climatology in the contiguous U.S.A. has been studied by Orgill and Sehmel (1976). Highest dust frequency (with visibility $< 11 \text{ km}$) 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. Other significant areas with lower 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 12.00 and 20.00 h local standard time. In the Great Plains, Hagen and Woodruff (1973) found that the average dust storm lasted 6.6 h and estimated the median dust concentration to be 4.83 mg m^{-3} at observer height. From visibility and windspeed data, approximate calculations showed that in the Great Plains during the 1950s, 221 Mt of dust were suspended annually and 70 Mt in the 1960s (Hagen and Woodruff, 1975).

These results suggest that particulate suspension from wind erosion exceeds that from all other sources (both natural and artificial) in the U.S.A.

Saltation

The characteristics of saltation (hopping) particles in wind have been described by several researchers (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 (or aggregates) lift off the surface (eject) and follow distinctive trajectories under the influence of air resistance and gravity. Such particles (100–500 μm) rise at fairly steep angles but are too large to be suspended by the flow. They return to the surface where they may abrade themselves or other aggregates on impact, or they may rebound or embed themselves and initiate movement of other particles. The bulk of total transport, roughly 50–80%, is by saltation. Saltating particles commonly rise < 120 cm; most rise < 30 cm.

Chepil (1945) reported lift-off angles of 75–90° and impact angles of 6–12° from the horizontal. White and Schulz (1977), using microbeads to study particle trajectories in a wind tunnel, reported an average lift-off angle of $50 \pm 20^\circ$ and average impact angle of $14 \pm 3^\circ$. Also, 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 (1945) found appreciable particle rotation of 200–1000 r.p.s. From changes in light reflections of microbeads on film, White and Schulz (1977) estimated spinning rates of 115–500 r.p.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 solution 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, saltating particles usually end up in a fence row, ditch, trap strip, windbreak or edge of a vegetated area downwind.

Surface creep

Sand-sized soil particles or aggregates 500–1000 μm in diameter, too large to leave the surface in ordinary erosive winds, are pushed, rolled and driven by the impacts of spinning particles in saltation. In high winds, the whole surface appears to be creeping slowly forward. The rippling of wind-blown sand has been attributed to unevenness in surface creep flow (Bagnold, 1941). Report-

edly, surface creep constitutes 7–25% of total transport (Bagnold, 1941; Chepil, 1945; Horikawa and Shen, 1960). Creep appears nearly passive in the erosion process, but creep-sized aggregates may abrade into the size range of saltation and suspension and, thus, shift modes of transport. Creep aggregates seldom move far from their points of origin.

Abrasion

The percentage of erodible soil ($< \sim 1000 \mu\text{m}$) in the surface layer is highly correlated with the mass of soil removable from that surface in wind-tunnel tests (Chepil, 1958). On long fields, the amount of soil that passes from a control volume on the soil surface increases nearly linearly with field length (Chepil, 1957b). Such a result implies abrasive breakdown of both erodible and non-erodible aggregates. Indeed, on long, erosion-susceptible fields, the total amount of soil that can be lost is usually several times the amount of erodible material initially present at the surface. Thus, resistance to abrasive breakdown of surface aggregates is important in wind erosion. However, until recently the effects of abrasion in the erosion process have been neglected and the physics of soil aggregate abrasions was largely unknown.

An abrasion susceptibility term (w) can be defined as the mass of material abraded from target aggregates per unit mass of impacting abrader. To determine how various factors effect w , large soil aggregates (50–100 mm in diameter) have been abraded with sand particles and soil aggregates using a calibrated nozzle (Hagen, 1984). The results show that

$$w = f(V_a, \alpha, d_a, S_t, S_a, \rho_a) \quad (4)$$

where V_a is the average velocity of the abrader, α is the abrader impact angle with the surface plane, d_a is the average abrader diameter, S_t and S_a are dry mechanical stabilities of the target aggregates and abrading aggregates, respectively, and ρ_a is abrader density. Major factors influencing w were V_a , α and S_t . A second report (Hagen and Lyles, 1985) showed that the amount of fine suspendible particles ($< 53 \mu\text{m}$) was markedly due to breakdown of saltating aggregates on impact rather than breakdown of target aggregates. These fine particles were enriched 3.1, 2.3, 1.7 and 1.9 times in nitrogen, phosphorus, potassium and organic matter, respectively, compared with their parent aggregates.

Sorting

Unless surface-layer aggregates or particles are homogeneous in physical properties (size, shape, density), which is highly unlikely in agricultural soils, sorting will occur during erosion. Sorting here refers to the selective removal during erosion of aggregates or particles because various sizes move at different

mass-flow rates. The sorting process over time removes the finer, nutrient-enriched materials, leaving behind those that are coarser and less fertile.

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

$$U_{*t} = A (\alpha g d_p)^{\frac{1}{2}} \quad (5)$$

where α is immersed density ratio, $(\rho_p - \rho)/\rho$, 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 0.08–0.12, perhaps as great as 0.2 without saltation 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 air kinematic viscosity. In Wood's summary of previous research, A ranged from 0.08 to 0.22 for $R_f > 0.7$; however, most values were between 0.08 and 0.12. Iversen et al. (1976) extended Bagnold's equation to include interparticle forces (caused by 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 eqn. (5). Some question remains concerning what value to use for d_p in erodible materials of wide particle-size range.

There is a growing tendency to extend the threshold concept to include boundary geometry, i.e. non-erodible aggregates, vegetation and microtopography, thus making it dependent on many physical dimensions and geometric patterns. I suggest that the term 'surface threshold' be used to separate this case from the more common 'particle' threshold, which is limited to flow and particle properties.

PARTICLE FLOW RATES

When considering particle flow rates, we must distinguish between cases in which all the particles are erodible ($< 1000 \mu\text{m}$) and cases in which there are mixtures of erodible particles and non-erodible elements, e.g. aggregates or clods $> 1000 \mu\text{m}$ and vegetative materials.

All erodible particles

The simplest case involves all erodible particles where several equations have been developed to estimate soil (sand) flow rates per unit width under specific soil and wind conditions. Most equations empirically developed from wind tun-

nel data relate flow rate to surface-shear stress or friction velocity of the wind and erodibility characteristics of the soil.

The functional form of those equations is

$$q = f[(\text{soil properties})^a, (\text{flow properties})^b] \quad (6)$$

where q is particle flow rate (mass per unit width per unit time) and a and b are constants. The soil properties term may include particle or aggregate size, density and shape; the flow properties term may include mean windspeed 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. The constant b commonly equals 3, which shows the well-known cubic relationship between friction velocity and particle flow rate and further supports the frequent use of U_* to indicate the wind's capability to erode soil. Specific flow rate 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). Flow rates vary considerably among the equations because of different values found for the constants and for coefficients introduced in explicit equations. All those particle flow-rate equations assume steady or stationary mean flow. Fan and Disrud (1977) suggested that mean wind velocities are generally not stationary and that steady flow equations may under-predict actual erosion.

Mixtures of erodible particles and non-erodible elements

Field soils seldom contain only erodible size particles; mixtures of erodible and non-erodible 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 non-erodible fractions, field roughness, vegetation and soil water content. No reliable equations have been developed to express field soil flow rates for short time periods or for single erosive windstorms. Recently, Cole (1984) has outlined procedures for determining field wind erosion rates from wind-tunnel-derived functions but no integratable functions for field surface conditions were presented. Currently, an effort is underway by the United States Department of Agriculture, Agricultural Research Service, to develop model(s) to predict the mass of soil lost from a field (area) during a specified time interval, including single erosion events.

The protective role of non-erodible elements in the erosion process is clearly due to absorption of part of the total wind drag, reducing the drag on erodible particles (Lyles et al., 1974). As erodible material is removed, exposing additional non-erodible material (e.g. clods), the drag on erodible particles decreases to the threshold value where erosion ceases and the soil is stabilized. Stabilizing agricultural fields by non-erodible elements is complicated by vari-

ations in speed, direction and duration of winds plus generation of erodible-size particles from larger aggregates by abrasion.

From wind tunnel studies, Lyles and Allison (1976) developed a regression equation that predicts the degree of protection provided by standing crop residues and non-erodible soil aggregates

$$\left(\frac{U_*}{U_{*t}}\right)_s = 1.638 + 17.044 \frac{NA_s}{A_t} - 0.117 \frac{L_y}{L_x} + [(1.0236)^c - 1] \quad (7)$$

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 as follows: 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; c , percentage of dry soil aggregates $> 1000 \mu\text{m}$ in diameter (limited to $\leq 50\%$). Research is needed to test eqn. (7) under field conditions.

PRINCIPLES AND GENERAL STRATEGIES OF CONTROL

In general, wind erosion can be decreased only by reducing wind forces on erodible aggregates and particles or by creating aggregates or surfaces more resistant to wind forces. Recognition that various components of the erosion process may be altered has led to the identification of four specific principles of wind erosion control (Woodruff et al., 1977), listed below.

- (1) Establish and maintain vegetation or vegetative residues.
- (2) Produce or bring to the soil surface non-erodible 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 (see Tibke, 1988; Ticknor, 1988) vary in space (from place to place) and may change over time along with cropping and management systems. Because saltation initiates and sustains suspension flow and initiates and drives the creep flow, control practices should focus on ways to stop or reduce saltation flow.

The four principles of control plus a factor for climate were used by Woodruff and Siddoway (1965) to develop the now well-known wind erosion equation, which predicts potential average annual erosion rates

$$E = f(I, K, C, L, V) \quad (8)$$

where E is the potential average annual soil loss in mass per unit area, I is the

soil erodibility index, K is the soil-ridge-roughness factor, C is the climatic factor, L is the unsheltered, weighted travel distance of the wind across a field and V is the equivalent vegetative cover. Briefly, the soil erodibility index, I , is the potential average soil loss from a wide, unsheltered, isolated field with bare, smooth, non-crusted surface based on climatic conditions near Garden City, Kansas. It is related to the percentage of non-erodible soil clods or aggregates in the surface layer ($> \sim 1000 \mu\text{m}$, actually $840 \mu\text{m}$). For convenience, the United States Department of Agriculture, Soil Conservation Service uses wind erodibility groups based primarily on soil texture as a guide for selecting I factors.

The soil-ridge-roughness factor, K , is a measure of the effect of ridges on erosion rates compared to a smooth (non-ridged) surface. It is determined from the height and spacing of ridges created by tillage implements.

The climatic factor C characterizes the erosive potential of climate (wind-speed, precipitation and air temperature) at a particular location compared to that at Garden City, Kansas, which is assigned an annual value of 100% based on long-term data.

The L factor recognizes that more wind erosion occurs on large fields than on small fields. It also considers that winds at a given site usually have a prevailing direction and that more winds blow in the prevailing direction in some places than in others. Cole et al. (1983) discussed two 'weighting' methods that have been used to determine L , and Skidmore (1986) has proposed another.

The vegetative cover factor, V , is based on the quantity, kind and orientation of vegetation or vegetative residues. All vegetative materials must be converted to an equivalent small-grain standard before use in the equation. Data for converting various plant materials to their small-grain equivalent are given by Woodruff and Siddoway (1965), Lyles and Allison (1980, 1981) and Armbrust and Lyles (1985). Generally, the V factor recognizes that, on a dry-weight basis, plants with small stalks are more effective in controlling erosion than plants with larger stalks, that standing plants (or residues) are more effective than flattened and that fine-leaved crops like wheat and other grasses provide a high degree of erosion control per unit weight.

Procedures have been developed for applying eqn. (8) to periods shorter than 1 year, which involves partitioning erosion amounts over time with erosive wind-energy distribution as the criterion (Bondy et al., 1980). Regardless of modifications, the wind erosion equation was developed to estimate the potential erosion from a particular field or to estimate field conditions necessary to reduce potential erosion to tolerable amounts and has been used extensively for design and evaluation of control systems (practices). In recent years (1977, 1982), it has also been used in the U.S.A. for making national erosion inventories.

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