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Wind erosion refers to the detachment, transport, and deposition of sediment by wind. It is a dynamic, physical process where loose, dry, bare soils are transported by strong winds. Geomorphologists and other earth scientists usually consider wind erosion as a specific subdiscipline of a more broad study of aeolian (also spelled eolian) processes. The term *aeolian* is derived from the Greek god Aeolus, the keeper of the winds, so aeolian processes refer to effects produced by the force of the wind interacting with surface features. Although aeolian research spans a wide range of topics, which may even include research on other planets, in this chapter we will limit our focus to erosion of soils by the wind on the Earth's surface, and more specifically on crop land and range land.

The movement of sediment by wind has been occurring for many eons, as demonstrated by aeolian cross-bedding seen in wind-blown sands of ancient sandstone bedrock. Loess deposits are ubiquitous accumulations of aeolian sediments of silt, and smaller amounts of clay and sand, derived from wind-blown glacial outwash deposits or from deserts or playa lakes. Large dune fields and sand seas around the world provide further evidence of current and past aeolian environments (Fig. 14|1). Fixed or stable dunes are no longer active in the current climate but were active sand seas or dune fields in the past.

Scientists have long been interested in the direct and indirect effects of wind erosion. The earliest publication relating to aeolian processes was written by a Flemish astronomer, Godefroy Wendelin, in 1646 (Stout et al., 2009). Wendelin's paper (Wendelin, 1646) described the purple rain of Brussels that we now recognize as wet deposition of African windblown dust. Charles Darwin collected dust over the Atlantic Ocean that had fallen during his voyages on the HMS Beagle (Darwin, 1845). Recent analysis of this dust indicated it originated from the western Sahara and molecular-microbiological methods demonstrated the presence of many viable microbes even today (Gorbushina et al., 2007).

Wind erosion is a soil degrading process that affects more than 500 million hectares of land worldwide and creates between 500 and 5000 Tg of fugitive dust annually (Grini et al., 2003). Perhaps the most memorable period of recent sustained wind erosion in the United States was the Dust Bowl era from about 1931 through 1939 (Baumhardt, 2003). During this period, wind erosion of rangeland and cropland reached an annual peak of 20 million hect-

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ares (Hurt, 1981), while the entire area affected encompassed almost 40 million hectares (Baumhardt, 2003).

Wind erosion has been estimated in the United States by the USDA-NRCS by means of a periodic Natural Resources Inventory (NRI). The NRI is a statistical survey of natural resource condition and trends on non-Federal U.S. land (USDA-NRCS, 2007). In 2003, the estimated erosion on cropland due to wind was 776 million tons per year. This represents a 7% reduction in erosion by wind estimated in a similar NRI compiled in 1997. In comparison, the amount of erosion on cropland due to water in 2003 was 971 million tons for the same year. These estimates are based on a longitudinal sample survey based on scientific sampling principles. Figure 14|2 shows the areas of total wind and water erosion. Although we are making progress in reducing wind erosion, it continues to be a national and international problem. In this chapter we will review the onsite and offsite effects of wind erosion, as well as details of the wind erosion process, prediction, and control measures.

Effects of Wind Erosion

Onsite Effects

The movement of large quantities of aeolian sediment as suspended windblown dust is clearly evident still today and produces dramatic on-site and off-site effects. Wind erosion winnows the finer, more chemically active components of the soil, especially nutrients affecting plant growth (Lyles, 1975; Sterk et al., 1996; Stetler et al., 1994; Van Pelt and Zobeck, 2007; Zobeck and Fryrear, 1986a,b). Other chemical species are lost in disproportionate amounts and unique chemical species such as anthropogenic radioisotopes may be used to estimate historic erosion rates in affected soils (Van Pelt et al., 2007). In addition to soil fertility degradation, the disproportionate loss of soil organic carbon (Van Pelt and Zobeck, 2007) and soil fines may affect soil water infiltration and holding capacity, further affecting soil productivity in semiarid regions.

In addition to soil loss from valuable agronomic systems and fragile natural ecosystems, wind erosion creates several other problems of great economic impact. In source fields, moving soil particles sandblast crop plants and can seriously damage a seedling stand (Armbrust, 1968; Fryrear and Downes, 1975; Skidmore, 1966). This damage often results in replanting decisions for producers (Fryrear, 1973). For example, cotton (*Gossypium hirsutum* L.) lint and kenaf (*Hibiscus cannibinus* L.) yields were reduced 40% and sorghum [*Sorghum bicolor* (L.) Moench] yields were reduced up to 58% in a study of a severely wind-eroded field in west Texas (Zobeck and Bilbro, 2001). In addition, for certain crops and certain growth stages, sandblast injury may result in increased rates of growth in surviving plants (Fig. 14|3; Baker, 2007). According to Farmer (1993), deposition of wind-blown soils on crops decreases their yield and hinders their processing.

Visibility reductions that may happen suddenly can result in a hazard to transportation and commerce on highways close to source fields. Dust storms often reduce visibility to less than 10 meters, causing numerous traffic accidents and deaths in developed countries. Numerous accidents have been attributed directly to wind-driven sand and dust (Skidmore, 1994). In one dust storm near Lubbock, TX in June 2006, 21 vehicles were involved in six different accidents sending 23 people to local hospitals, with one death reported (Blackburn, 2006).

Deposition of wind-driven sand along field margins, especially along weedy fence lines and in drainage ditches, results in costly, recurring maintenance tasks for landowners and government authorities (Fig. 14|4). Recent research indicates that most wind-eroded soil is deposited very close to the source field (Hagen et al., 2007). Wind-eroded soil that is not deposited along field margins enters the suspension mode and may be lofted tens to thousands of meters in altitude in the turbulent boundary layer (Gillette et al., 1997; Chen and Fryrear, 1996; Zobeck and Van Pelt, 2006). Dust that leaves the field and is transported significant distances is termed *fugitive dust*.

Offsite Effects

Fugitive dust impacts environmental, animal, and human health, as well as industry, transportation, and commerce for tens to hundreds of kilometers downwind. The

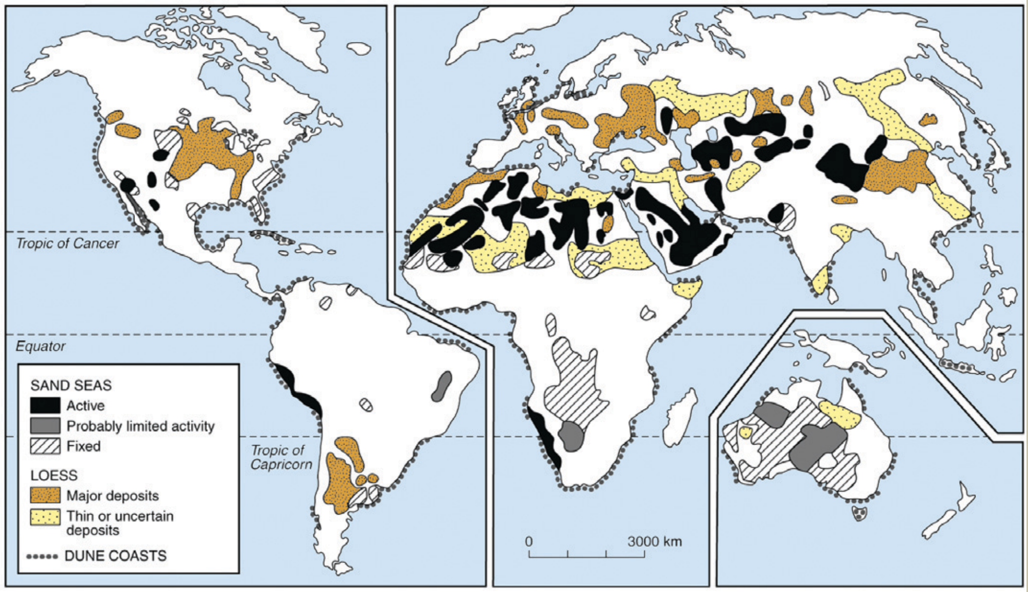

Fig. 14|1. Location of sand seas, loess, and dune coast deposits (with permission from Thomas and Wiggs, 2008).

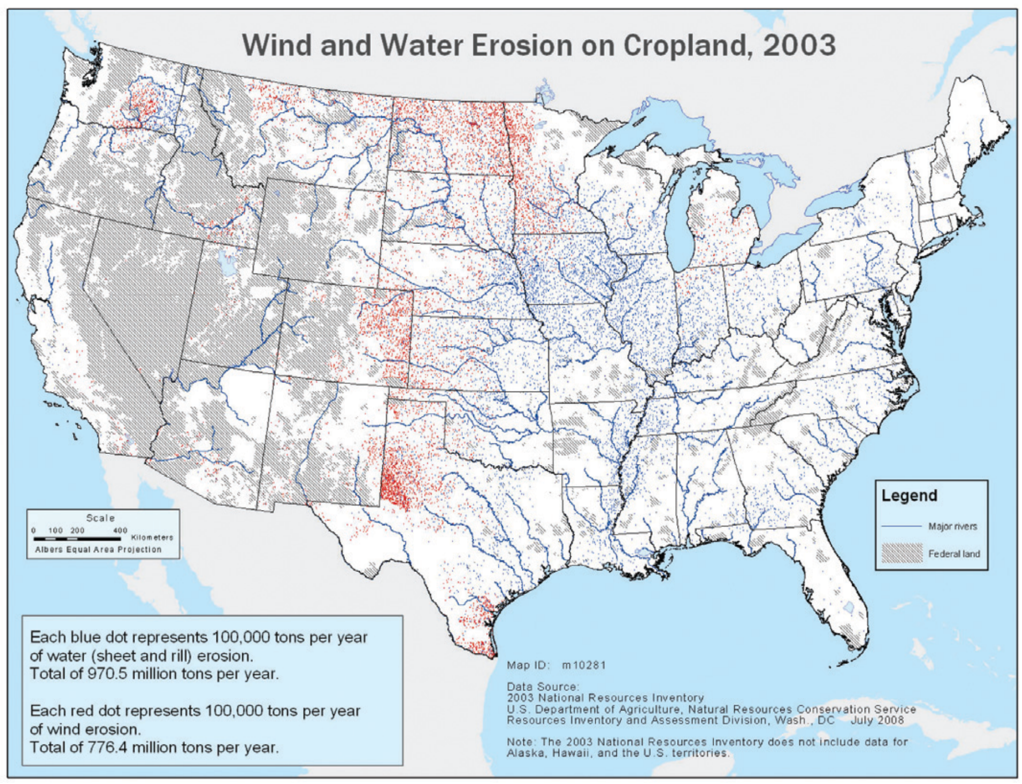

Fig. 14|2. Estimated average annual wind and water erosion on cropland in the United States estimated in the 2003 National Resources Inventory (USDA-NRCS, 2007).

Fig. 14|3. Examples of cotton plant damage after exposure to sand abrasion for 0, 5, 10, 20, 30, and 40 min, left to right (from Fig. 1 in Baker, 2007).

Fig. 14|4. Drainage ditch filled with sand following a severe dust storm in west Texas.

Clean Air Act, amended in 1990, required the U.S. Environmental Protection Agency to establish National Ambient Air Quality Standards (NAAQS) and set limits on airborne pollutants, including fine particulate matter. The standards were designed to protect public health and welfare, including protection against decreased visibility, damage to animals, corps, vegetation, and building (USEPA, 2008). Wind erosion has been reported as a major cause of noncompliance of the NAAQS within the Columbia Basin, Washington (Saxton, 1995).

Wind currents and circulation patterns are capable of carrying smaller diameters of fugitive dust between continents. Dust from the Saharan Desert in Africa has been documented to have fallen in Europe (Goudie, 1978), South America (Talbot et al., 1990), the Caribbean Sea (Delany et al., 1967), the North Atlantic Ocean (Prospero, 1996), and to the interior of North America, a distance of more than 9000 km from the source region (Gatz and Prospero, 1996). Similarly, dust from the deserts of northern China has been documented in Korea (Chung et al., 2003), Japan (Lee et al., 2003), North America (Shao, 2000), Alaska (Rahn et al., 1981), and Hawaii (Braaten and Cahill, 1986). Mineralogical analysis has indicated that the majority of dust deposited in the glaciers of Greenland originates from eastern Asia (Svensson et al., 2000).

Dust is an important agent for transporting soil parent material (Gile and Grossman, 1979; Reynolds et al., 2006), plant nutrients, trace metals (Van Pelt and Zobeck, 2007), soil biota (Delany et al., 1967), and toxic anthropogenics (Larney et al., 1999) between ecosystems and watersheds. The great loess deposits in various areas of the world (Fig. 14|1) are from aeolian deposition (Tsoar and Pye, 1987) and deposition of lesser amounts of aeolian sediments may affect the properties of soils weathered from bedrock or fluvial sediments (Rabenhorst et al., 1984). The mineralogy, chemical, and biotic characteristics of soil dust are determined by the surface from which it is entrained (Reheis and Kihl, 1995). Microbiological exudates such as fatty-acid methyl esters (Kennedy, 1998) or enzyme activities and arylsulfatase proteins (Acosta-Martinez and Zobeck, 2004) may be used to identify the probable source area of a given dust outbreak. Pathogenic microbes may also be transported on dust and affect distant ecosystems and human health (Leathers, 1981). Some agronomic ecosystems depend on the inputs from deposited dust (Sterk et al., 1996). Iron fertilization and resultant blooms of algae in the oceans has been documented and may result in increased carbon dioxide sink and photosynthetic production of oxygen (Mackie and Hunter, 2007). However, deposition of nutrient-rich dust in freshwater lakes and over terrestrial watersheds may result in undesirable algal blooms in freshwater bodies.

During transport, dust may enter into numerous chemical reactions and catalyze reactions of anthropogenic particulates in the atmosphere. Calcium carbonate is a common soil constituent in semiarid and arid regions and thus is a common constituent of soil dust (Gile and Grossman, 1979). Calcium carbonate originating from the Owens Lake dry lakebed is partially converted to calcium sulfate before it is deposited in the Los Angeles basin (Reheis and Kihl, 1995). Acid rain is a problem worldwide, but is partially ameliorated in regions where carbonate-rich dust interacts with the acid species in the clouds or in the soils of the affected watershed (Litaor, 1987; Trochkine et al., 2003). In regions distant from anthropogenic oxides of sulfur and nitrogen, the carbonates in dust may make normally mildly acidic rainwater alkaline (Zhang et al., 2003).

Calcium carbonate particles that have been modified by reactions with atmospheric acids are more hygroscopic and tend to form more effective condensation nuclei (Krueger et al., 2004). These wetted soil aerosols may attract and absorb gases and other aerosols from the adjacent atmosphere, allowing for rainout and effectively cleaning the atmosphere. Humic acid coatings on soil dust are highly attractive to hydrophobic organic species in the atmosphere (Chiou, 1989). The catalytic effect of humic acid coated soil dust in the atmosphere is enhanced at high relative humidities as the hydrophobic nature of the humic acid is overcome and the particles adsorb a thin layer of water (Brooks et al., 2004). Additionally, Miller et al. (1989) showed humus in the presence of sunlight to be an effective catalyst, creating highly reactive free radicals of oxygen that are instrumental in the oxidation of organic pollutants.

Wind Erosion Mechanics

Wind erosion occurs as the wind interacts with the soil surface to cause detachment (termed *entrainment*), transport, and finally deposition of soil particles. Detachment occurs as the wind exerts drag and lift forces to overcome the gravitational and cohesive forces that hold particles to the soil surface (Toy et al., 2002). Detachment also occurs as rolling or bouncing particles cause other particles to be released by impacts or abrasion (Fig. 14|5). The wind velocity at which sediment begins to move is termed the *threshold wind velocity*. Winds are considered erosive when they reach a speed of about 6 m s^{-1} (13 mph) at 0.3 m (1 ft) above the soil

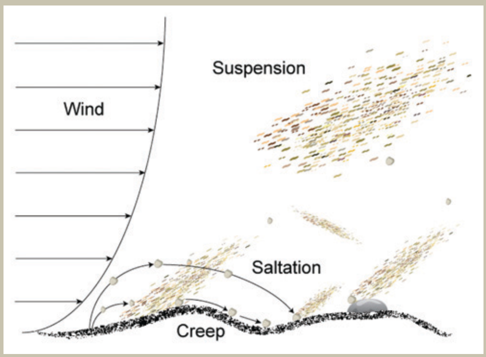

Fig. 14|5. Modes of particle transport due to the force of the wind on the surface. Length of wind arrows indicates relative strength of the wind.

surface or about 8 m s^{-1} (18 mph) at 9 m (30 ft) above the surface (USDA-NRCS, 2002).

After the wind exceeds the threshold wind velocity, soil particles or small stable aggregates begin to move in three primary ways, or modes, of transport: creep, saltation, and suspension (Fig. 14|5). Particles or soil aggregates in creep mode are about 0.5 to 1 mm in diameter and roll or scoot along the soil surface, propelled by the direct force of the wind or when bouncing (termed *saltating*) particles strike them. Individual saltating particles or soil aggregates are about 0.1 to 0.5 mm in diameter and move by bouncing along the soil surface, rarely exceeding heights of a few meters. These particles may be directly lifted off the soil surface by the force of the wind or be ejected from the soil surface as other saltating particles dislodge them on impact. As these saltating particles bounce along the soil surface they dislodge even more saltating particles, creating an avalanching effect. Saltating sediment may cause abrasion as particles bounce or collide with other sediment or the crusted soil surface, or they may become lodged in the soil surface.

Suspended sediment is generally less than 0.1 mm in diameter. Although some suspension-sized sediment is present in the soil, it is less susceptible to direct entrainment by the force of the wind. Suspended material is mainly created as saltating sediment abrades larger aggregates or strikes the soil surface in a process similar to sandblasting. Saltating particles collide with the surface with a force that is a function of their mass and velocity. However, recent studies in the Columbia Basin of Washington suggest direct emission of suspension is possible in some silty soils (Kjelgaard et al., 2004). Although particles less than 0.1 mm can be suspended, particles larger than about 0.02 mm diameter are unlikely to travel greater than 30 km from the source, settling back to the surface quite quickly when the turbulence associated with strong winds abates (Pye, 1987). The finer suspended sediments are carried up by turbulent eddies and, as mentioned above, may travel thousands of kilometers before settling back to the surface. In contrast, creep and saltating sediments are usually redeposited within or near the source field.

Wind

As the wind interacts with the Earth's surface, the surface exerts a drag on the wind, reducing the wind velocity nearer the surface (Fig. 14|5). During strong wind events, the boundary layer near the surface is usually statically neutral and the vertical profile of wind speed may be described by a well-known semilogarithmic equation of the form:

$$u(z) = \frac{u^*}{k} \ln\left(\frac{z}{z_0}\right) \quad [1]$$

where $u(z)$ is the wind speed at height z, u^* is the friction (or shear) velocity, k is the von Kármán constant (0.4), and z_0 is the aerodynamic roughness height. The friction velocity is a measure of the shear stress on the surface and has been used in predictive models as the driving force for wind erosion. It is indicative of the atmospheric turbulence and is proportional to the slope of the wind velocity profile when the height is represented on a logarithmic scale (Fig. 14|6). The aerodynamic roughness height refers to the theoretical height at which the wind speed near the surface falls to zero and depends on the characteristics of the surface. Numerous studies have found that z_0 is approximately equal to 1/30 the height of the roughness

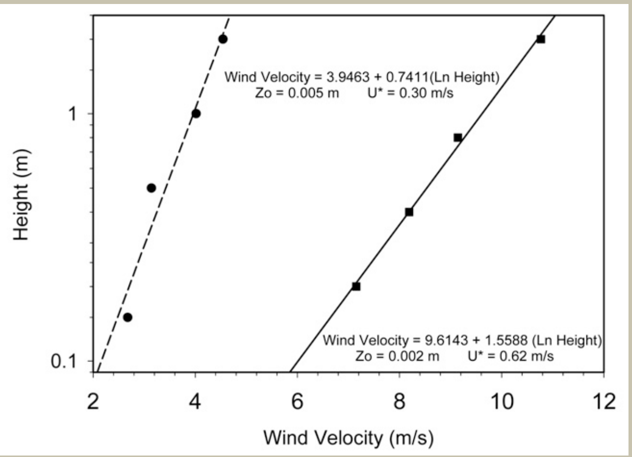

Fig. 14|6. Wind profiles above two soil surfaces. U* is the friction velocity and Zo is the aerodynamic roughness height.

elements. In vegetated surfaces, z_0 may vary with wind speed as the vegetation bends in the wind. In effect, the aerodynamic roughness is a representation of the capacity of the surface for absorbing momentum and is also an important quantity in wind erosion studies (Shao, 2000). In practice, if we plot wind speed on the y axis and the logarithm of the height on the x axis, we normally obtain a straight line with the slope u^*/k and the intercept $(u^*/k)\ln(z_0)$.

Although the wind provides the energy to drive wind erosion, the characteristics and condition of the soil surface will ultimately control whether or not erosion occurs and its extent. In the next section we will explore how soil surface conditions affect wind erosion.

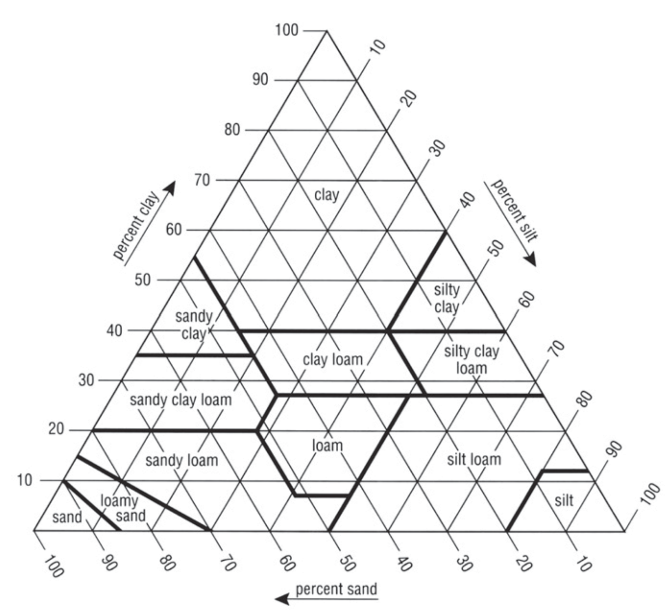

Fig. 14|7. Soil textural triangle.

Soil Surface Conditions

Soils have been described as having intrinsic or inherent soil properties that change very slowly and dynamic or temporal properties that vary through time. Dynamic soil properties may change very rapidly in response to weather factors, tillage, or other management and include properties such as bulk density and dry aggregate size distribution. Examples of inherent soil properties are soil texture, organic matter content, and mineralogy.

Soil Texture

The USDA-NRCS has established soil textural classes (Fig. 14|7) based on specific proportions of sand, silt, and clay contained in a sample (USDA-NRCS, 1993). Soil texture is one of the primary soil properties affecting soil susceptibility to wind erosion (also called wind erodibility). The USDA-NRCS has classified the wind erodibility of soils according to the soil texture and calcium carbonate content (Table 14|1). In general, coarse soils such as sands are more erodible than finer-textured soils such as clay loam soils. Calcareous soils tend to have a higher erodibility than noncalcareous soils. Calcareous soils contain enough calcium carbonate to cause effervescence with the application of dilute acid. Soil texture and calcium carbonate content are inherent soil properties that change very slowly with time. Even so, this does not mean that it is impossible to change the texture. For example, soil scientists in west Texas have found that the surfaces of soils that have undergone wind erosion for a long period of time are now coarser than when originally mapped several decades ago. The relative increase in sand content over time has been caused by the winnowing of finer particles out of the soil by wind erosion.

Soil Moisture

Surface soil moisture content is an extremely important variable controlling both the entrainment (erodibility) and transport of sediment by wind (Nickling, 1994). Wind tunnel experiments have shown that soil moisture content clearly affects the wind threshold friction velocity at which

Table 14|1. Relation of soil texture and soil erodibility.†

Soil texture‡	Predominant soil texture class of surface layer	Wind erodibility group (WEG)	Soil erodibility index (I)§
			Mg ha^{-1} yr^{-1}
C	Very fine sand, fine sand, sand, or coarse sand	1	694¶
			560
			493
			403
			358
C	Loamy very fine sand, loamy fine sand, loamy sand, loamy coarse sand, or sapric organic soil materials	2	300
C	Very fine sandy loam, fine sandy loam, sandy loam, or coarse sandy loam	3	193
F	Clay, silty clay, noncalcareous clay loam, or silty clay loam with more than 35% clay	4	193
M	Calcareous loam and silt loam or calcareous clay loam and silty clay loam	4L	193
M	Noncalcareous loam and silt loam with less than 20% clay, or sandy clay loam, sandy clay, and hemic organic soil materials	5	125
M	Noncalcareous loam and silt loam with more than 20% clay, or noncalcareous clay loam with less than 35% clay	6	108
M	Silt, noncalcareous silty clay loam with less than 35% clay, and fibric organic soil material	7	85
–	Soils not susceptible to wind erosion due to coarse surface fragments or wetness	8	–

† Adapted from the USDA National Agronomy Manual (USDA-NRCS, 2002).

‡ Soil texture: C, coarse; M, medium; F, fine.

§ The erodibility index is based on the relationship of dry soil aggregates greater than 0.84 mm to potential soil erosion.

¶ The I factors for WEG1 vary from 358 Mg ha^{-1} yr^{-1} for coarse sands to 694 Mg ha^{-1} yr^{-1} for very fine sands. For coarse sands gravel, use a low figure. For very fine sand without gravel, use a higher value. When unsure, use an I value of 493 Mg ha^{-1} yr^{-1}.

particles begin to move (Belly, 1964; Bisal and Hsieh, 1966; Chepil, 1956). Early studies by Chepil (1956) suggested that soil erodibility by wind was about the same for soil that was oven-dried or air-dried when moisture content did not exceed one-third of the 15-atmosphere percentage (–1500 J kg^{-1} matric potential). Beyond this range of moisture a distinct decrease in erodibility was observed. Above about 5% gravimetric moisture content, sand-sized material is inherently resistant to entrainment by most natural winds (Nickling, 1994). More recent studies have related the change in threshold friction velocity with soil water tension, derived from capillary force equations that consider the capillary forces developed at interparticle contacts surrounded by water (McKenna-Neuman and Nickling, 1989). The erodibility of soil by wind is so sensitive to the effects of moisture that even differences in relative air humidity modify the particle threshold wind velocity (McKenna-Neuman and Sanderson, 2008; Ravi et al. (2006a,b). A wind tunnel study of sand, sandy loam, and clay soils showed that the threshold friction velocity decreases with increasing values of relative humidity for values between 40 and 65%, while above and below this range the threshold fiction velocity increases with air humidity (Ravi et al., 2006b).

Surface Roughness

The aerodynamic roughness length is determined from the wind profile and is the height above the surface at which the mean wind speed becomes zero. This empirically derived value is related to roughness elements on the soil surface (e.g., clods, rocks, vegetation), as well as the surface microtopography or microrelief. The effects of vegetation on roughness will be discussed later in the chapter. Soil surface microrelief is a dynamic soil property that may change rapidly due to management or weather factors.

In tilled agricultural soils, tillage produces an oriented roughness (or ridges) parallel to the direction of tillage caused by pulling the tillage tool through the soil. In addition, a random roughness is produced by the random orientation of soil aggregates or clods on the surface. Research has shown that wind erosion is sensitive to the effects of both random and oriented roughness (Fryrear, 1984). The roughness caused by tillage will modify the wind profile to change z_0 and also protect the soil surface from the effects of abrading particles. In general, the aerodynamic roughness length increases as the size of the clods or ridges increases. The amount of change is related to the size and spacing of the ridges and clods. In addition, the effects of the roughness produced by tillage on wind erosion will depend on wind direction when ridges are present. When erosive winds blow perpendicular to a bare soil surface, ridges will act to physically protect a fraction of the soil surface, as illustrated in Fig. 14|8. The ridges have little effect when the wind blows in the direction parallel to tillage. A microrelief index called the cumulative shelter angle distribution (CSAD) is used to estimate the fraction of the tilled soil surface susceptible to abrasion by saltating grains (Potter and Zobeck, 1990). The CSAD has been shown to be sensitive to tillage tools, rainfall, and wind direction.

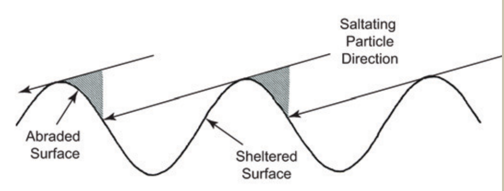

Fig. 14|8. Schematic representation of a ridged field. Part of the field is sheltered from abrasion by saltating grains.

Aggregate Properties

Soil aggregates or peds are naturally occurring structural units composed of primary soil particles (USDA-NRCS, 1993). They are formed as a consequence of natural soil development. Cohesion within the aggregates is greater than the cohesion among adjacent aggregates. Thus, they are formed when stress causes the soil to rupture under predetermined planes of weakness. Soil clods are similar to aggregates, but soil forming processes have exerted very little or no control on the boundaries of clods. They are produced by tillage or other soil manipulations that cause the soil to rupture and break apart, and they may include pieces of aggregates. Following tillage, the soil surface typically contains clods and aggregates with a wide range of sizes.

Dry aggregate size distribution refers to the relative amounts of air-dry aggregates and clods, on a mass basis by size class, present on the soil surface (Zobeck, 1991b). A rotary sieve is used to determine the dry aggregate size distribution (Chepil, 1962). Wind erosion is related to the amount of aggregates >0.84 mm in diameter, called *nonerodible aggregates*. Table 14|2 lists the soil wind erodibility, also called the *I* value, as a function of the percentage of nonerodible aggregates. The dry aggregate size distribution

Table 14|2. Soil wind erodibility as determined by percentage of nonerodible soil (>0.84 mm diameter).†

	Units									
	0	1	2	3	4	5	6	7	8	9
Tens‡	Mg ha^{-1} yr^{-1}									
0	–	694	560	493	437	403	381	358	336	314
10	300	293	287	280	271	262	253	244	237	228
20	220	213	206	202	197	193	186	181	177	170
30	166	161	159	155	150	146	141	139	134	130
40	125	121	116	114	112	108	105	101	96	92
50	85	81	74	69	65	60	56	54	52	49
60	47	45	43	40	38	36	36	34	31	29
70	27	25	22	18	16	13	9	7	7	4
80	4	–	–	–	–	–	–	–	–	–

† Adapted from the USDA National Agronomy Manual (USDA-NRCS, 2002).

‡ Columns and rows represent the percentage of nonerodible aggregates. For example, to find 33% go to the nonerodible aggregates tens row at 30 and units column at 3 (30 + 3 = 33) to find 155 Mg ha^{-1} yr^{-1}.

is commonly expressed as the geometric mean and geometric standard deviation derived from a lognormal distribution or as the shape and scale parameters of a Weibull distribution (Zobeck et al., 2003a). The Weibull distribution has been shown to be more accurate and precise in describing dry aggregate size distributions for tilled soils (Zobeck et al., 2003a).

Dry aggregate stability refers to the resistance of soil aggregates to breakdown from physical forces. It is a measure of the bonding strength of the bonding agents within aggregates (Skidmore and Powers, 1982). The physical forces causing aggregate breakdown may occur as a result of tillage but may also include the physical forces caused by the impact of saltating grains. The dry aggregate stability of bulk samples of tilled soils has been determined by repeated sieving using a rotary sieve (Chepil, 1958). In this case, the dry aggregate stability is calculated as the weight of the particles or aggregates greater than 0.84 mm in diameter after the second sieving divided by the weight of the particles or aggregates greater than 0.84 mm in diameter after the first sieving. The stability of individual soil aggregates is determined by measuring the force needed crush the aggregate to a known endpoint (Hagen et al., 1995). In this case, the energy needed to crush an aggregate approximately 15 mm in diameter is called the *crushing energy*. The stability of dry aggregates has been shown to be a dominant predictor of soil erosion from surface abrasion (Hagen, 1991a). Skidmore and Layton (1992) found that aggregate clay content and water content at the wilt point (–1500 J kg^{-1} matric potential) are good predictors of mean aggregate stability.

Surface Crusting

Surface crusting refers to a relatively thin consolidated soil surface layer or seal that is more compact and cohesive than the material immediately below it. When crusts are formed, particles are bound together and less susceptible to abrasion by blowing soils than the less stable material below the crust. Under natural conditions, crusts form from a variety of physical, chemical, and biological processes (Neave and Rayburg, 2007). Details of the interparticle forces contributing to the cohesion of crusts have been described by (Ishizuka et al., 2008). In rangelands, biological cryptogamic crusts may be particularly effective in stabilizing the soil by binding small particles into larger, nonerodible aggregates and protecting the soil from wind erosion (Eldridge and Greene, 1994; Leys and Eldridge, 1998).

In cropland soils, primary tillage acts to mix and loosen the surface. Rainfall is the primary agent that can create the crust or seal after tillage. The effects of rainfall on soil crust development have been studied for many years. Most studies have shown that the strength of the crust to withstand abrading particles, known as *crust stability*, is related to the soil properties and rainfall rate or energy used to create the crust (Zobeck, 1991b). The dislodgement of surface particles was shown to decrease with increasing crust strength in a laboratory study using artificially created crusts (Rice et al., 1996). In a wind tunnel study of 14 crusted soils that included a wide range of soils textures (from loamy sand to clay and one organic soil), loose, unconsolidated soil was on average about 40 times as erodible as crusts created using simulated rainfall at an intensity of 25 mm h^{-1} and 70 times as erodible as crusts created using simulated at an intensity of 64 mm h^{-1} (Zobeck, 1991a). In addition, the crust abrasion was positively correlated with the sand content and cation exchange capacity/clay ratio.

In sandy soils, loose, unconsolidated erodible material may be left exposed on the crust after rainfall. This loose, erodible material (LEM) is highly susceptible to wind erosion. If LEM is not present on a crusted surface, wind erosion generally will not occur. The LEM acts as projectiles or bullets as they bounce or saltate, abrading (sandblasting) the surface. The mass of LEM on crusted soils is affected by inherent soil properties, management, and climatic factors (Zobeck, 1991a). Rainfall simulation studies have shown that the logarithm of LEM was related to sand content, sampling location in relation to tillage ridges, and rainfall (Potter, 1990). Sandy soils tend to have much more LEM on the crust than finer textured soils. In Potter's (1990) rainfall simulation study of five soils ranging in texture from fine sandy loam to clay, the fine sandy loam soil had about 30 times the amount of LEM as the clay soil tested.

Stones, stable nonerodible aggregates, vegetation, and other nonerodible materials will physically protect or armor the soil surface from the direct force of the wind and from abrading sand particles. The effect of this protection is related to the amount of nonerodible material on the soil surface, as described in Fig. 14|9. The soil loss ratio described in Fig. 14|9 is the ratio of the erosion observed for the protected soil divided by the erosion on bare, unprotected soil. Covering the soil with 20% nonerodible material reduced soil loss by 57% and covering the soil with 50% nonerodible material reduced soil loss 95% (Fryrear, 1985).

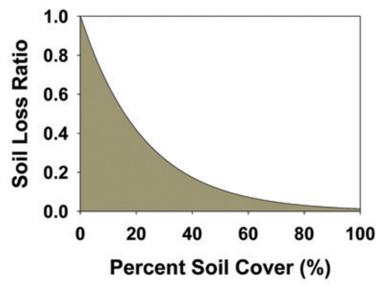

Fig. 14|9. Soil loss ratio as a function of soil cover by nonerodible elements.

Wind Erosion Models for Cropland

A wide variety of models have been developed to predict sediment entrainment and transport at scales ranging from small plots to global. A review of many of these models has been presented by Zobeck et al. (2003b). In the U.S., the Wind Erosion Equation (WEQ) (Woodruff and Siddoway, 1965) and the Wind Erosion Prediction System (WEPS) (Hagen, 1991b) have been the principle models used to predict wind erosion on cropland.

Wind Erosion Equation

The wind erosion equation (WEQ) was designed to predict long-term average annual soil erosion by wind based on a specific set of climatic and field conditions. It can also be used to predict erosion for specific time periods when using the appropriate factors in the equation. Details about WEQ are found in the USDA-NRCS *National Agronomy Manual* (USDA-NRCS, 2002). The WEQ is determined using the following equation:

$$E = f(IKCLV) \qquad [2]$$

where E is the estimated annual soil loss (mass/unit area/time period), f indicates a nonlinear functional relationship among the variables, I is the soil erodibility index (mass/unit area), K is the soil surface roughness factor, C is the climatic factor, L is the unsheltered distance, and V is the vegetative cover factor.

In practice, the soil erodibility index (I) is determined by the surface soil texture and assigned a value as indicated in Table 14|1. The I value is the potential annual wind erosion for an isolated, unsheltered, level, wide, bare, smooth, loose, and noncrusted soil at a location where the climatic factor is equal to 100. The other individual factors are then multiplied, after their determination using measured values, nomographs, or other methods, as indicated in Eq. [2]. The K value adjusts the I value for tillage-induced random and oriented roughness. The C value adjusts after consideration of wind speed and potential evapotranspiration. This factor is expressed as a percentage of the C factor for Garden City, KS, which has a C factor of 100. The L factor considers the unprotected distance across the field along the direction of the prevailing wind direction. The V factor considers the kind, amount, and orientation of vegetation on the surface. The USDA-NRCS *National Agronomy Manual* (USDA-NRCS, 2002) has detailed instructions on the use of WEQ, and a spreadsheet version of WEQ for use in the United States was provided by (Sporcic et al., 1998). The wind erosion equation is currently used by USDA-NRCS for conservation planning and assessing soil erosion by wind for the USDA-NRCS National Resources Inventory.

Wind Erosion Prediction System

Although the WEQ has been used successfully for many years, it has several limitations. The WEQ predicts average annual erosion by summing erosion predicted for specific periods of time (i.e., 2-wk periods) but does not predict for daily events. The WEQ does not account for changes in many temporal surface features affecting wind erosion, such as changes in surface roughness and aggregation, and it does not account for the two-dimensional nature of fields. The climatic factor of WEQ is related to a standard location and does not stochastically model weather conditions. Many details in crop and soil management were not considered in the development of WEQ. In an attempt to resolve these issues, USDA-ARS embarked on the development of a new wind erosion prediction technology, the Wind Erosion Prediction System (WEPS).

The WEPS is a process-based, daily time-step, computer model that predicts soil erosion by wind by simulating the fundamental processes involved (Hagen, 1991b). The current version of WEPS (1.0) is designed to provide the user with a simple tool for inputting initial field and management conditions, calculating soil loss, and displaying simple or detailed outputs for designing erosion control systems. The WEPS is the computer implementation of a science model that simulates the processes involved with the wind erosion process. The science model is composed of the following major submodels (USDA-ARS, 1996):

- Weather—Uses historical statistical information of a wide variety of meteorological variables with stochastic techniques to determine the likelihood of various variables needed to drive processes in other submodels.
- Hydrology—Uses inputs from other submodels to compute water content in the various soil layers and at the soil–atmosphere interface throughout the simulation period.
- Management—Models the primary human-initiated actions that affect the susceptibility to wind erosion. These include all cultural practices applied to the field, such as tillage, planting, harvesting, and irrigation.
- Soil—Simulates soil temporal properties that affect the susceptibility to wind erosion on a daily basis in response to driving processes such as rainfall, tillage, and others.
- Crop—Calculates daily production of masses of roots, leaves, stems, reproductive organs, and leaf and stem areas.
- Decomposition—Simulates the decrease in crop residue biomass due to microbial activity and includes standing, surface, buried, and root biomass pools.
- Erosion—Uses parameters supplied by other submodels to simulate the process of soil movement. The submodel periodically updates any changes in the soil surface caused by soil movement and outputs estimates of soil loss or deposition from the simulation region.

In practice, users define a simulation region (field) and input management and soils information using a graphical user interface. The program includes a variety of databases from which information about barriers, soils, management, crop and decomposition, and climate is extracted. For example, after the location of the field is identified, the interface selects the appropriate weather station for which historical data are used to simulate weather parameters. The soils data are selected from a soils database supplied by the NRCS Soil Survey. WEPS provides a wide variety of output, including soil loss as saltation/creep, suspension, and particulate matter less than 10 μm (PM_{10}). By varying inputs, particularly those related to management, the user can easily evaluate various erosion control alternatives.

Initial tests of WEPS indicate that it produces reasonably good estimates of soil loss due to wind when compared with measured results. For example, soil loss measurements from 46 storms events from eroding fields in six states had reasonable agreement ($R^2 = 0.71$) with erosion simulated by WEPS (Hagen, 2004). A comparison of WEPS with measured soil loss in Germany showed excellent agreement ($R^2 > 0.9$) between measured and simulated soil loss (Funk et al., 2004). The USDA-NRCS is currently testing the WEPS for application in the United States as a replacement for WEQ. Detailed information about obtaining WEPS is available at http://www.weru.ksu.edu/weps/wepshome.html (verified 7 Dec. 2010).

Sensitivity analyses have shown that predictive models are very sensitive to soil surface conditions. The WEQ is particularly sensitive to soil texture, surface roughness, and residue. The sensitivity of WEQ to texture is so great that investigators have suggested using texture-adjusted *I* factors to calibrate the model for local soil conditions (Van Pelt and Zobeck, 2004). The WEPS has been shown to be sensitive to soil surface conditions, including soil surface wetness, dry aggregate stability, oriented roughness, and residue management (Hagen et al., 1999). In a separate evaluation of the WEPS for 46 wind events at six North American locations, Hagen (2004) noted that model inaccuracies may be due to average soil parameter values being used or time-dependent wind erosion-induced changes in soil erodibility after the date of soil sampling. In a very detailed sensitivity analysis, Feng and Sharratt (2005) reported that the WEPS was most sensitive to changes in biomass flat cover, near-surface soil water content, ridge height, and other management-related parameters, including crust cover and random roughness. The sensitivity of these models to management-related parameters is indicative of the profound effects that land management has on wind erosion.

Management Effects on Wind Erosion

Native Vegetation Communities

It is widely held that land management has a profound effect on erosion of the soil surface by wind. In humid and subhumid climates, the canopy of native vegetation communities is generally sufficient to prevent erosive wind energy from reaching the soil surface. For most forest ecosystems, native grasslands, and managed pastures, the literature is largely lacking reports of observed wind erosion. In semiarid and arid climates, however, native plant communities do not fully protect the soil surface from the erosive forces of wind. Semiarid ecosystems including grasslands, shrubland, savanna, woodland, and forests are all susceptible to wind erosion, especially when disturbed.

In a series of studies of several communities ranging from relatively undisturbed ponderosa pine (*Pinus ponderosa* P. Lawson & C. Lawson) forests to desert shrublands dominated by mesquite (*Prosopis glandulosa* Torr.). Breshears et al. (2008) measured sediment transport rates ranging from 0.17 to 27.4 g m^{-2} d^{-1}, respectively. These sites ranged from 75 to 0% woody canopy cover and from 98 to 38% herbaceous ground cover. For disturbed sites in the same or similar locations, they measured sediment transport rates of 1.1 to 6002 g m^{-2} d^{-1}, with total canopy coverage of from 72 to 0%, respectively. They further concluded that sediment transport may be inherently greater for shrublands and that sparse shrublands have a greater influence on the wind profile by channeling the wind and increasing turbulence. Forests, dense shrublands, and grasslands protect the soil surface more evenly and tend to produce a "skimming flow" profile by the wind. Natural areas devoid of vegetation due to ephemeral flooding or water diversion are also susceptible to wind erosion (Pelletier, 2006). Ephemeral playa lakes, when dry, constitute locally and regionally important areas of dust generation (Prospero et al., 2002).

The effects of vegetation cover on wind erosion and soil loss have been investigated in a desert grassland of southern New Mexico (Li et al., 2007). The authors concluded that as lateral cover, a function of plant number density and vertical dimension, drops below 9%, wind erosion increases dramatically. Anthropogenic disturbance of desert grasslands by mechanical means or overgrazing often results in a sparse shrubland subclimax. The process begins by exposing the soil surface to wind (Sharifi et al., 1999; Liu and Wang, 2007) and becomes exacerbated by sandblasting of the remaining vegetation (Okin et al., 2001). The result is an anisotropic pattern of shrub vegetation (McGlynn and Okin, 2006), which leads to further degradation of the landscape by erosion of the soil surface in the bare alleys (Okin and Gillette, 2001) and deposition of the finer, more nutrient-rich samples in and under the canopies of the shrub patches. This redistribution of soil fines leads to heterogeneity of soil texture and infiltration rates that further influences the distribution of vegetation (Ravi et al., 2007). The recognition of the dependence of wind erosion on the distribution of vegetation in these disturbed communities has led to the development of a stochastically based model of wind erosion in sparsely vegetated communities (Okin, 2005, 2008).

Burning of native vegetation communities also increases the susceptibility of the soil to erosion by wind and subsequent degradation. Burning increases the erosion hazard primarily by removing the vegetation, which both slows the wind near the surface and prevents the re-entrainment of deposited sediments (Stout, 2006). The heat of the fire may also alter physical properties of the soil and affect the soil stability (Whicker et al., 2008). Fire has been credited with increasing the water repellency of soils and thus increasing their erosion susceptibility by maintaining a dry surface and modifying the surface soil threshold friction velocity (Ravi et al., 2006a). The effects of fire that increase the susceptibility of the soil to wind erosion may be short-lived, however, as vegetation grows back and protects the surface.

Cropped Ground

The development of land for production agriculture is often accomplished by total native vegetation removal and at least some smoothing of the land surface, leading to the increased susceptibility of the soil to wind erosion. Conventional cropland tillage practices that lead to the increased susceptibility of the surface to wind erosion and dust emissions include plowing, delaying primary tillage, leveling beds, planting, weeding, fertilizing, cutting and baling, spraying, and burning. Management methods that are used to control wind erosion on cropland include planting windbreaks to alter wind flow patterns, strip cropping, planting cover crops before or after low residue crops, cross-wind strips, vegetation barriers, retaining plant residue after harvest, stabilizing the surfaces using water or applied chemicals, and tilling the field to bury erodible particles, to increase roughness by increasing the percentage of nonerodible aggregates on the surface, and to create bed patterns perpendicular to the predominant winds (Nordstrom and Hotta, 2004; USDA-NRCS, 2009).

Windbreaks and shelterbelts have been used to decrease the erosive force of the wind in many local settings. They are typically rows of trees and shrubs planted along the margins of the field or farmstead they are intended to protect, but they may also be fences, rock walls, or earth berms. Such barriers effectively decrease the wind speed for a distance of about 10 to 15 times their height downwind and about three times their height upwind (Oke, 1987). Due to the limitations of tree growth in many regions, especially semiarid regions, this distance rarely exceeds a few hundred meters downwind (Vigiak et al., 2003). Windbreaks and shelterbelts are not as common as they once were. Although there were approximately 65,000 km of them planted in the Great Plains of North America by the 1960s (Griffith, 1976), by the 1970s, many were dying or were being removed (Sorenson and Marotz, 1977).

Maintaining crop residues on the cropped ground is perhaps the most effective management solution for controlling wind erosion. The value of crop residues for controlling wind erosion has been recognized for at least six decades (Chepil, 1944). Residue protects the ground by offering elements that prevent saltating particles from cascading and by increasing the roughness height z_0 (Eq. [1]). The WEQ and other predictive models treat all crop residues, whether standing or flat on the ground as protection equivalent to flat small grain residues (Woodruff and Siddoway, 1965; Bilbro and Fryrear, 1985). Standing residues and growing crops provide greater protection than flat residues because they absorb much of the shear stress in the boundary layer (Skidmore, 1994). This displaces the effective roughness height, z_0, by a zero plane displacement height, d, which is a factor of the height, density, and stiffness of the vegetation (Oke, 1987). It may be approximated for a wide range of crops and trees by:

$$d = \frac{2}{3}h \qquad [3]$$

where h is the mean height of the standing crop or residue. The displacement height d is used to modify the logarithmic wind profile Eq. [1] over a crop under adiabatic atmospheric stability (Monteith, 1973):

$$u_z = \frac{u^*}{k}\ln\left(\frac{z-d}{z_0}\right) \qquad [4]$$

The effects of vegetation or other nonerodible material on the soil surface on soil loss is estimated using the soil loss ratio, an index calculated by dividing the amount of soil loss from a residue-protected soil surface by the loss from a similar bare surface (Fig. 14|9). The soil loss ratio decreases rapidly from 1.0 for a bare unprotected surface to a value of approximately 0.2, an inflection point characterized by 40% soil cover (Fryrear, 1985).

Another description of plant canopy or residue used by predictive models for standing vegetation is the plant silhouette through which the wind must pass. Bilbro and Fryrear (1985) observed a strong relationship between plant silhouette and the soil loss ratio. However, very sparse residue or other roughness element cover may actually increase soil loss by compressing airflow and creating localized super-critical wind velocities that exceed threshold (Sterk, 2000).

Tillage of cropped land tends to bury crop residues and thus diminish the protection provided. Tillage may also weaken aggregate stability by decreasing the soil organic carbon content (Fenton et al., 1999; Six et al., 1999) and thus increase the soil's intrinsic erodibility. Tillage is used to protect the soil surface by increasing the nonerodible fraction and the roughness of the surface (Fryrear, 1984). Oriented roughness elements, such as ridges created using tillage implements such as a lister, are very effective when oriented at angles greater than 13 degrees from the direction of the incident wind (Hagen and Armbrust, 1992). Nonoriented or random roughness is also used to lessen wind erosion by creating numerous nonerodible elements that provide shelter from abrading sand grains (Potter et al., 1990).

Best Management Practices for Controlling Wind Erosion

The best management practice (BMP) to prevent erosion is to prevent contact of wind with the soil surface by maintaining an effective cover of residue, such as a cover crop or carefully managed stubble. The emergence of no-till and conservation tillage practices has resulted in more effective post-harvest standing and flat residue over cropped ground. Advances in harvest equipment, such as finger or stripper headers on small grain combines, have also led to improvements in the post-harvest heights of standing residue. In semiarid regions that represent marginal dryland farming regions and with certain locally important crops such as cotton or sunflowers (*Helianthus annuus* L.), insufficient silhouette or flat residue may fail to protect the soil. In addition, cultural practices in some crops, such as tillage for insect control in cotton, can contribute to loss of standing and flat residue.

In areas where rainfall is not limiting, such as the Red River Valley in Minnesota, sugar beets are protected with a low rate of spring barley (40 kg/ha [0.75 bu/ac]) before planting. The low rate of cover is killed with herbicides after the beets are established (M. Sporcic, personal communication, 2009).

On bare soils or soils with limited crop residues, tillage remains a common BMP to prevent erosion. Raising beds perpendicular to the prevailing wind direction increases the aerodynamic roughness and provides regularly spaced roughness elements offering shelter angle to prevent cascading saltation. By creating a surface dominated by nonerodible aggregates, a random roughness is formed that offers the same protective shelter angle to prevent cascading saltation. In fragile soils with low dry aggregate stability, erosion may start in localized areas of the field or at the downwind end of a long, frequently traveled, unpaved road. In such locations, it may be necessary to use a snow fence or other barrier to encourage deposition and discourage saltation. Intense rainfall on soils with low wet aggregate stability often results in a smooth crusted soil surface with loose sand-sized material on the surface. The use of crust breaking and clod forming tillage implements such as a rotary hoe or a sand fighter is often used after spring thunderstorms to create random roughness to the field surface. Once the crop is established and the canopy covers a significant portion of the soil, tillage is only used to control weeds.

Planting annual crop barriers is another BMP for soils in limited rainfall areas. For example, 102-cm (40-inch) strips of weeping love grass could be planted at 30- to 91-m (100–300 ft) intervals perpendicular to the erosive wind direction, depending on the soil properties. The interval can be determined using erosion models such as those described in this chapter (M. Sporcic, personal communication, 2009).

Wind erosion is a natural process that has formed and continues to form landscapes in anthropogenically disturbed and undisturbed locations. It is unlikely to think we can completely control wind erosion in every case. In cases where wind erosion is difficult to control, we need to evaluate all of the onsite and offsite costs and effects to determine whether or not we are pursuing the most prudent management. In most cases, control measures must be applied where economic losses and health or environmental effects from wind-driven soil movement are likely.

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