Structure and Function of Chihuahuan Desert Ecosystem
The Jornada Basin Long-Term Ecological Research Site
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Eolian Processes on the Jornada Basin

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In arid and semiarid lands, soil erosion by wind is an important process that affects both the surface features and the biological potential of the ecosystem. The eolian flux of soil nutrients into or out of an ecosystem results in enrichment or impoverishment of its biological potential. In the Jornada Basin, wind erosion is the only significant mechanism for the net loss of soil materials because fluvial processes do not remove materials from the basin. Vigorous wind erosion leads to topographic changes, altering the growing conditions for plants and animals. Examples of such changes in topography are the formation of sand dunes or the removal of whole soil horizons. Our goal in this chapter is to describe the construction of a mathematical model for wind erosion and dust production for the Jornada Basin. The model attempts to answer the following questions:

1. Which soils are affected by wind erosion?
2. How does wind erosion occur on Jornada soils?
3. Does changing vegetation cover lead to a change in the source/sink relationship?
4. Is the Jornada a source or sink of eolian materials? If it is a source, what materials are lost?
5. How does wind erosion change the soil-forming process?

We will provide provisional answers for the questions and outline work that will more clearly define these answers.

Airborne dust has a significant residence time in the atmosphere and acts to modify the radiative properties of the atmosphere, mainly by back-scattering the incoming solar radiation (Andreae 1996). Changing land uses in arid and semiarid areas (e.g., overgrazing and cultivation) can drastically alter the dust emissions to the atmosphere (Tegen et al. 1996). The climatic effects of soil-derived dust were investigated in an experiment in central Asia (Golitsyn and Gillette 1993). Using measured size distributions for emitted dust (Sviridenkov et al. 1993) and various real and imaginary indices of refraction (Sokolik et al. 1993), Sokolik and Golitsyn (1993) calculated climatic effects. Atmospheric dust decreased the total radiative balance of the underlying surface and at the same time induced general warming of the underlying surface–atmosphere system due to a decrease in the system albedo over the arid zones. Because of the climatic effects of dust, it is important to understand the mechanisms that determine the flux of dust in arid and semiarid locations. We need to understand whether humans are having an effect on the global flux of dust and consequently on that part of climatic change caused by a change of atmospheric burden of dust.

Assessing dust emissions requires a framework for identifying the importance of various mechanisms. This is a complex task because wind erosion involves nonlinear and threshold processes. Because the interactions are nonlinear and highly interactive, mathematical or physical modeling is desirable to predict the consequences of land use and to reconstruct past conditions affecting wind erosion. Some erosion processes are governed by universal relationships so well established (Greeley and Iversen 1985) that there is no reason to verify them. One such process is threshold friction velocity (u_{*t}) for particles on smooth surfaces. Threshold friction velocity is a measure of the minimum wind force needed to sustain wind erosion; it is the minimum friction velocity (u_*) to sustain wind erosion. Friction velocity is defined in equation 9-1. For other parameters in the wind erosion model (table 9-1), it is highly desirable to verify their application in the Jornada Basin.

Table 9-1. Physical relationships needed for the wind erosion model.

1. Effect on u_{*t} of aerodynamic roughness height (z_0), vegetation cover, and size distribution of loose soil.
2. u_{*t} for crusts (biological and rain-physical).
3. Effect of soil moisture on u_{*t}.
4. Aggregation/disaggregation: destruction of soil crust by sandblasting.
5. Particle supply limitation effect.
6. Total particle mass flux (q) as a function of u_* and u_{*t}.
7. Owen effect (increase of z_0 with u_* above u_{*t}).
8. Ratio of vertical flux of dust to horizontal flux of coarse particles (F_a /q).
9. Effect of vegetation on wind erosion.

Measurements of Threshold Friction Velocity and Small-Scale Aerodynamic Roughness Height, z_0

The soils affected most by wind erosion are those having the lowest threshold friction velocities (u_{*t}). We tested Jornada soils for threshold friction velocity using a portable wind tunnel described by Gillette (1978). The tunnel has an open-floored test section so that a variable-speed turbulent boundary layer can be developed over a flat soil containing small-scale roughness elements, such as pebbles and small aggregates of soil.

The wind tunnel has a two-dimensional 5:1 contraction section with a honeycomb flow straightener and an expanding rectangular diffuser attached to the working section in a configuration similar to that of Wooding et al. (1973). In field studies at the Jornada, the working section was 231 cm^2 in cross-section and 2.4 m in length, and the wind tunnel was placed in areas free of vegetation. Wind data were obtained 20 cm from the end of the working section at the midpoint of the tunnel width and at eight different heights spaced approximately logarithmically apart from 2 mm above the surface to 10 cm. The Pitot tube anemometer was calibrated against the NCAR reference wind tunnel and was corrected for the air density change caused by elevation above sea level. Data for the wind profiles were fitted to the function for aerodynamically rough flow.

$$U = (u_*/k)\ln(z/z_0) \qquad (9\text{-}1)$$

Where k is von Karman's constant (set to 0.4), U is mean wind speed, z is height above the surface, u_* is friction velocity, and z_0 is aerodynamic roughness height.

The threshold wind speed for wind erosion was defined to be that speed at which we observed small but sustained movement of particles across the soil surface. After slowly increasing the wind to the threshold of particle motion, we measured two sets of wind speed profiles. The following threshold profiles were obtained for each site. (1) For crusted soils, we measured the threshold for loose particles on the surface and for the destruction of the crust. For clay-rich soils of playa and other low elevation soils where there were no loose particles on the surface, only the breakup of the soil surface was measured. (2) At sites without crusts, the threshold for loose surface particles was measured. (3) At all sites, soils were disturbed using either livestock hoof impact or one pass of a 3/4-ton truck moving at a speed of about 8 km/h. These disturbances always created loose particles on the surfaces. Measurements were made of the loose-particle threshold immediately following the disturbance. For each site, two replicates of the threshold measurements were obtained. Sites were chosen to be representative of generic soil classifications made for the Jornada Experimental Range (JER): sandy, silty, clayey, and gravelly. Rock (nonerodible) surfaces were not tested.

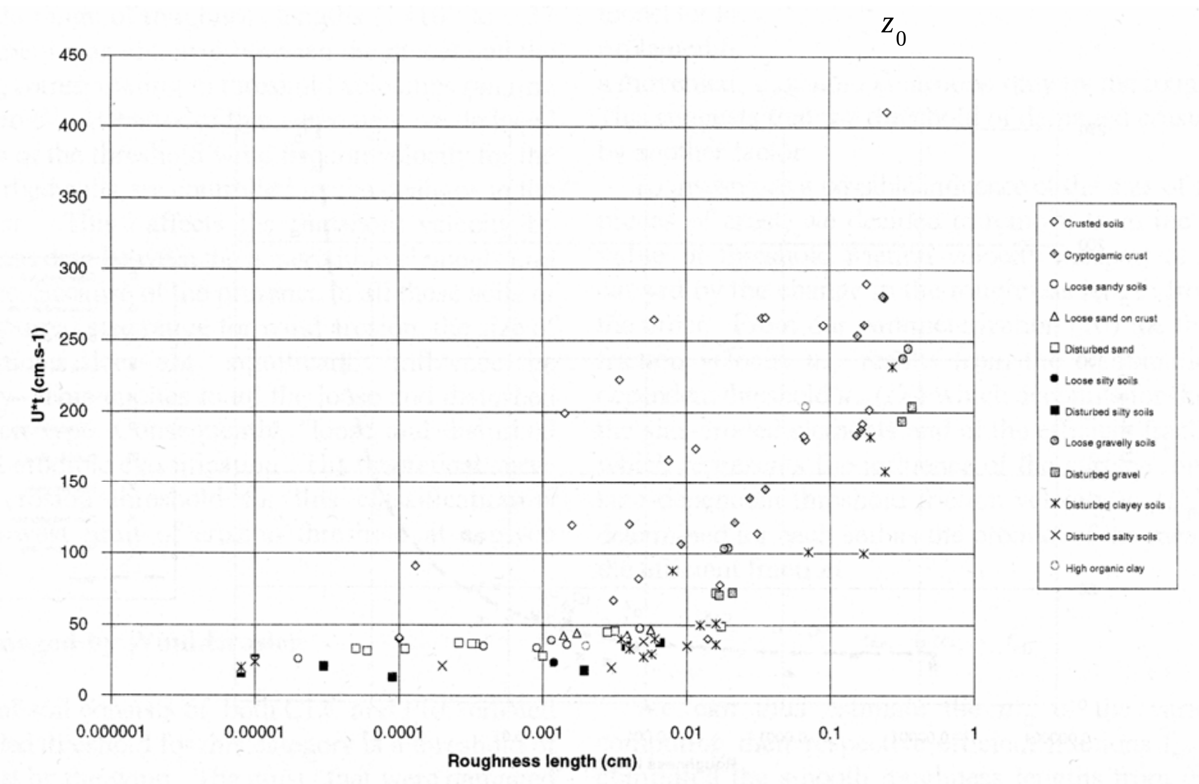

Fig. 9-1. Values of threshold friction velocity, u_{*t} , vs aerodynamic roughness length, z_0, (from Marticorena et al. 1997).

A plot of all results from wind-tunnel measurements of u_{*t} versus z_0 (aerodynamic roughness length) is shown in figure 9-1. The plot clearly shows that the lowest values of u_{*t} are found for disturbed soils and undisturbed, noncrusted sandy soils. The values for u_{*t} for sandy soils and disturbed soils (i.e., the lowest u_{*t}) versus z_0 values shown in figure

9.1 are shown in figure 9-2.

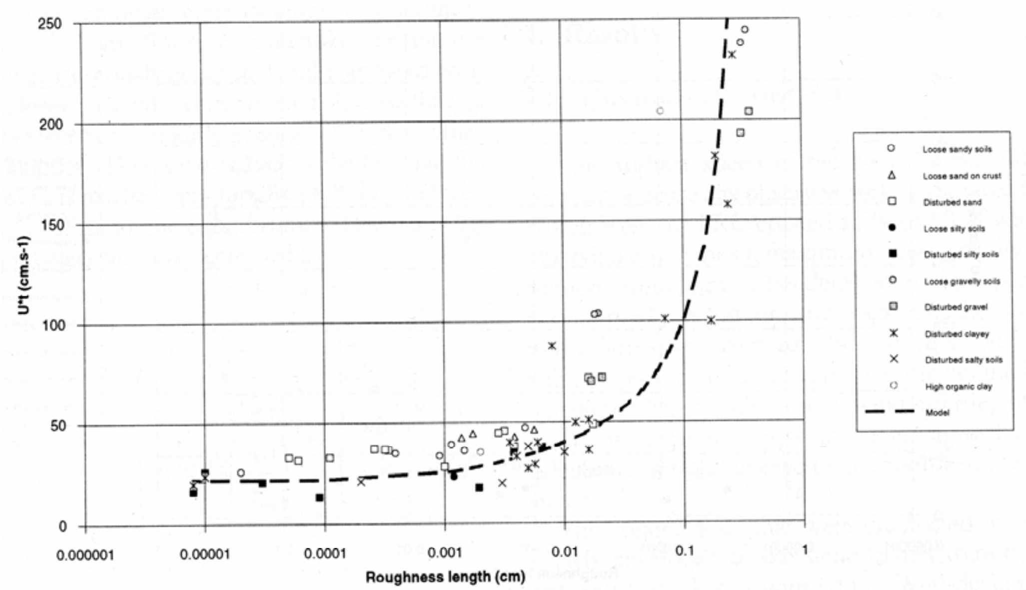

Fig. 9-2. Values of u_{*t} vs z_0 for undisturbed sandy soils and disturbed soils, not including gravels and cianobacteria lichen crusts. The curve is a model of u_{*t} vs z_0 assuming abundant particles 80 < d < 120 µm and a momentum partitioning scheme described by Marticorena et al. (1997)

The property that all the soils of figure 9-2 had in common was an abundant supply of loose surface particles of size 80 < d < 120 µm. This size is mobilized at a minimum threshold friction velocity of about 21 cm/s for smooth surfaces. Consequently, even though the soil size distribution varied from place to place, this variation did not affect the value of u_{*t} for a smooth surface. For disturbed and sandy soils, roughness of the surface z_0 controls u_{*t}. This roughness acts to absorb part of the momentum of the wind.

Cyanobacteria lichen crusts (CLC) and physical rain crusts (PRC) contain surface agglomerations of small soil particles bound by physical or biological agents. CLCs occur in open shrub and grass communities in arid and semiarid environments around the world as some combination of nonvascular microphytes (West 1990). Microphytes entangle and hold soil particles together, binding particles to form a crust that should resist wind displacement (Campbell 1979). For areas where vascular plants are sparse, CLCs help stabilize the soil against the wind (Williams et al. 1995). Belnap and Gillette (1997) showed that threshold velocities for sandy desert soils in southeastern Utah are significantly increased by CLCs, which protected soil otherwise at risk. PRCs are not biological in origin but bind particles together with silica, salt, gypsum, and clay, often forming crusts following individual rain events (Williams et al. 1995).

For both PRCs and CLCs, loose particles on the surface are usually much larger than size $80 < d < 120$ μm. Because the smooth threshold friction velocity of pieces of the loose surface crust was quite high compared to the size $80 < d < 120$ μm particles, both CLCs and PRCs had quite high threshold velocities and would erode only in unusually high winds. CLCs increase the threshold in two ways: The crusting roughens the surface, and the biological fibrous growth aggregates soil particles even after the crust is dry, as well as when the biological material is dead. Consequently, the CLCs are effective in the protection of the soil against wind erosion when undisturbed. When disturbed, the CLC loses some but not all of its protective qualities because disturbance smoothes the roughness and breaks the brittle aggregates. Threshold friction velocities for CLCs and PRCs at the Jornada were higher than the typical wind speeds recorded at the Jornada; consequently, areas of undisturbed CLCs and PRCs are areas without wind erosion (Belnap and Gillette 1998).

How Does Wind Erosion Occur on the Jornada Soils?

To estimate wind erosion at the Jornada, various physical relationships need to be developed or verified. At a study site (~ 1 ha) near the center of the JER, vegetation was cleared and a permanent array of instruments installed to measure erosion of a typical Jornada sandy soil subject to disturbance by grazing. The site can be expected to represent the most erodible of the Jornada soils.

The study site is roughly semicircular, having a diameter of > 100 m. Three meteorological towers of 2 m height were located on a line parallel to the dominant direction for wind erosion, southwest. Towers were located 50 m (west), 80 m (middle), and 110 m (east) from the southwestern edge of the clearing. Instrumentation on each of the three towers was as follows: wind speed at 0.2, 0.5, 1.0, and 2.0 m heights; air temperature at 0.2 m and 2.0 m heights; particle collectors at 0.05, 0.1, 0.3, 0.5, 0.6, and 1 m heights; and fast-response particle mass flux sensors at 0.05, 0.1, 0.2, and 0.5 m heights. In addition, markers were set in the soil so that increases or decreases in the height of the crust surface could be measured.

Data on the following relationships were collected:

- Owen effect (increase of z_0 with u_* above u_{*t}),
- q (total particle mass flux) as a function of u_* and u_{*t} including particle supply limitation,
- u_{*t} change by soil moisture,
- aggregation/disaggregation: destruction of soil crust by sandblasting, and
- ratio of vertical flux of dust to horizontal flux of coarse particles (F_a/q).

Owen Effect

The Owen effect is a feedback mechanism by which airborne sediment increases the drag coefficient of the surface. Before sand-sized particles are injected into the air (saltation), the air contains momentum that is transported by turbulent eddy transfer. During saltation, the sand grains interact with the air and transfer a part of the wind momentum to the ground. This occurs because the saltating particles strike the ground in a parabolic trajectory (implicit in the definition of saltation). These particles carry momentum absorbed at heights near the tops of their trajectories; this momentum transport is more efficient than that by air eddy transport. Consequently, momentum from greater heights in the air layer is required to replace the particle-transported momentum. This sequence of momentum transport is very similar to that caused by an increase of aerodynamic roughness length of the surface (z_0). The result is the formation of an internal boundary layer with a higher value of z_0. The increased u_* caused by saltating particles grows upward in the wind profile toward the height of the preexisting boundary layer. The Owen theory specifies that an increased z_0 will match the effect on the wind profile of the saltating particles. The effect is most easily detected as an increase of the ratio of friction velocity u_* to mean wind *U*. The Owen effect causes an amplification of sediment movement in a steady wind stream caused by nonhomogenous threshold friction velocity by locally increasing the drag coefficient. For a region having a uniform wind, the Owen effect can cause friction velocities to increase beyond those for areas that are not eroding.

Total Particle Mass Flux as a Function of u_* and u_{*t}, Including Particle Supply Limitation

Although a generalized function for soil particle flux has been given by Iversen and White (1982), it is worthwhile to verify the formula with data from the Jornada. Figure 9-3 shows the response of four Sensit detectors at the middle tower at the Jornada test site on October 22, 1995.

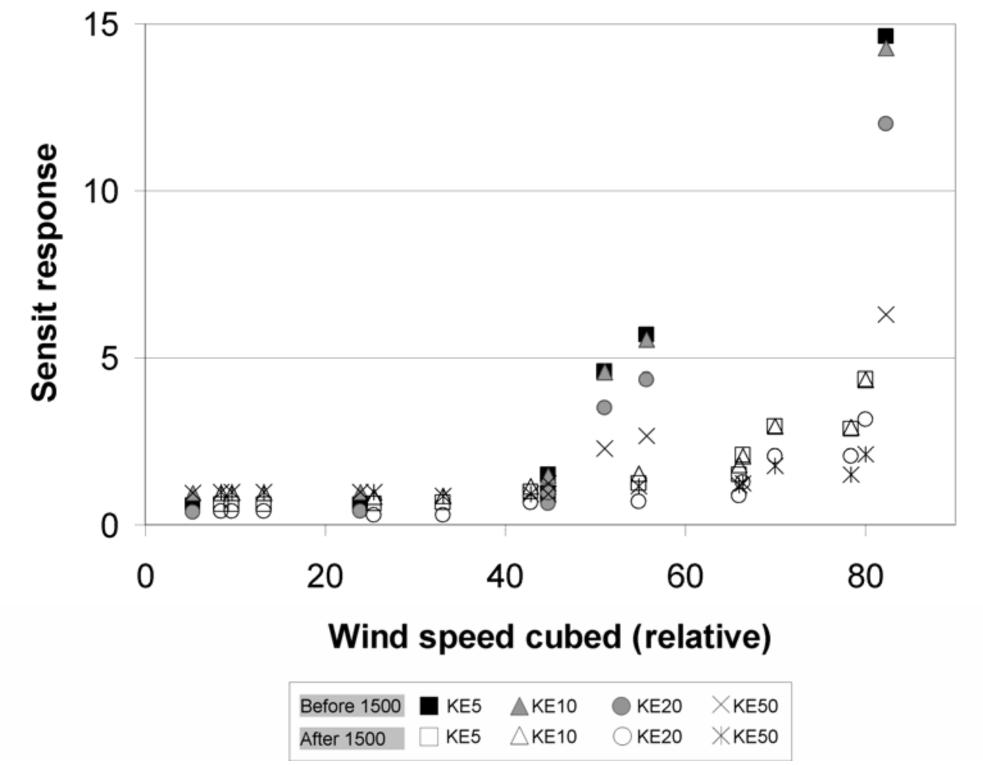

Fig. 9-3. Thirty-minute mass flux at a cleared, sandy soil site recorded at the middle tower vs u_*^3 for October 22, 1995, before and after 1500 h. The mass flux shows a linear relationship with the cube of friction velocity until 1500 h then decreases. Interpreted this as a depletion of available particle supply at 1500 h—particle supply limitation.

The data are categorized as "before 1500" (3 P.M.) and "after 1500." Sensit response is proportional to mass flux and these data were taken at 5-, 10-, 20-, and 50-cm heights. The before 1500 data show that the mass flux increases with the cube of wind speed in approximate agreement with the formula of Iversen and White (1982).

The mass flux data after 1500 show a decrease of slope. Data obtained before 1500 represent the particle flux from a layer of loose material; those after 1500 represent the new lower flux rate reflecting crust material that has a small mass of loose particles.

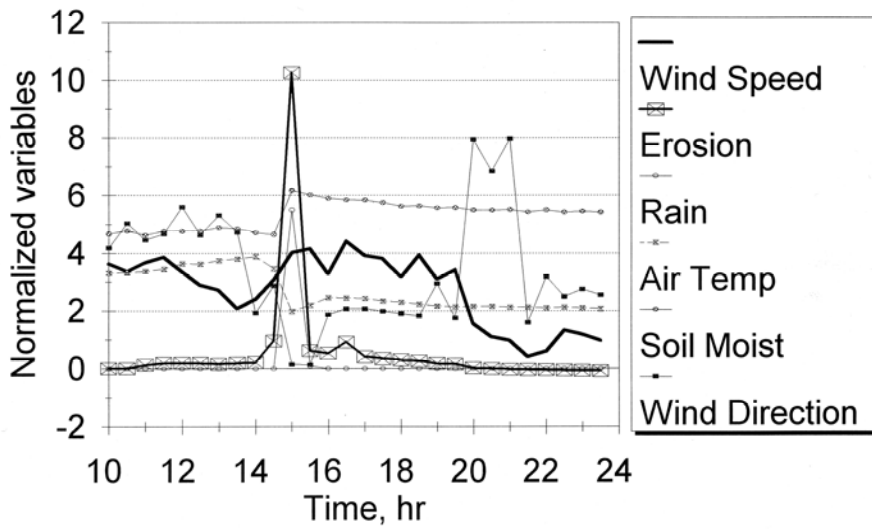

Fig. 9-4. Simultaneous half-hour mean measurements at the west vegetation-cleared site on August 7, 1996, of wind speed at 2 m, sand flux at 10 cm height, rainfall, air temperature at 20 cm height, soil moisture at a depth of 6 cm, and wind direction. All measurements are in relative units to show a qualitative picture of wind erosion conditions.

Consistent with this interpretation were observations that the sample site had a thin, loose cover of sand with several large bare patches of crust. The sharp decrease of the slope of mass flux from one constant value to another lower value probably coincided with the removal of a loose particle layer, leaving behind the supply-limited crust.

Soil Moisture Effect on Threshold Friction Velocity

Soil moisture can increase the threshold friction velocity of a soil (Chepil 1956; Bisal and Hsieh 1966; Saleh and Fryrear 1995). McKenna-Neuman and Nickling (1994) showed that sand grains are held together by the capillary effect of soil moisture. Qualitative observations of the effect of soil moisture on u_{*t} at the Jornada are shown in figure 9-4 for August 7, 1996.

Following a drop in air temperature and shift in wind direction at 1500 h, erosion flux increases just before the arrival of rain. Following the rain, soil moisture at 6 cm increases quickly. Although the wind reaches a higher value at 1630 than at 1500, there is greater mass flux at 1500. At 1830 h, the mass flux is zero even though the wind speed is about the same as it was at 1500 (the time of maximum wind erosion). This shows that threshold friction velocity increases with moistening of the soil. However, the figure also shows that the soil dries out quickly, so the effect of soil moisture is short-lived in this desert location.

Crust Formation/Disaggregation

The largest cause of the temporal and spatial variability of threshold friction velocities is the aggregation and disaggregation of soils that can change both u_{*ts} (threshold friction velocity for smooth surfaces) and z_0. Aggregation is a mechanism by which individual particles in the surface sediment become effectively larger particles or crusts. Aggregation or crust formation usually occurs with the drying of a moistened surface. Drying is a complicated process, and the aggregation that occurs is affected by the composition of the soil, evaporation rate, and temperature. Destruction of the crust and surface aggregates depends on sandblasting, freeze-thaw cycles, formation of crystals, and temperature of the sediment. Observations of crust destruction on the Jornada site are shown in figure 9-5. This figure shows the abrasion of the Jornada crust versus total sand passage over the surface. Both the vertical flux of kinetic energy on the crust and the sand passage are proportional to the cube of the wind speed. Because soil aggregates are known to abrade as a function of the kinetic energy of the sandblasting (Hagen et al. 1992), the linear relationship of the crust abrasion to mass transport *q* is not surprising.

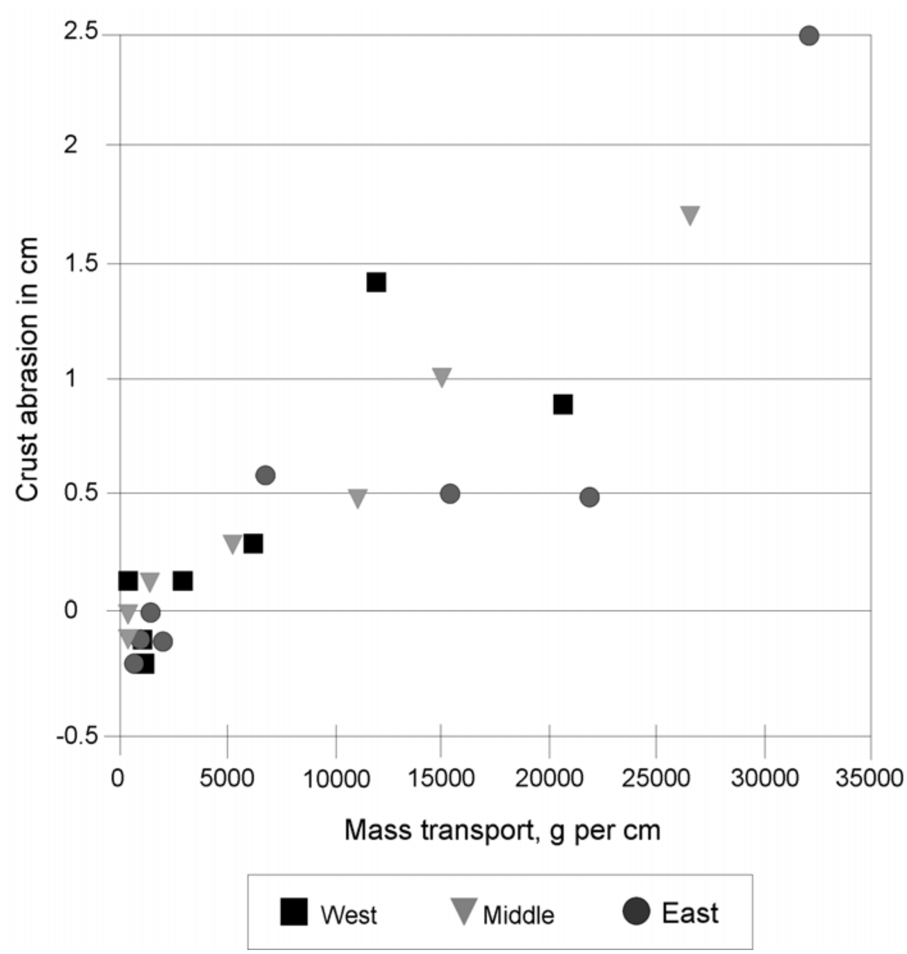

Fig. 9-5. Crust abrasion plotted against total sand movement for the three towers at the vegetation-cleared site for August 1995 to March 1996. The data suggest a linear relationship between crust abrasion and sand movement.

Relation of Vertical Flux of PM_{10} Dust to Total Particle Mass Flux

Shao et al. (1993) modeled the vertical flux (F_a) of wind-eroded particles smaller than 10 μm (PM_{10}) and related that flux to the saltation-particle mass flux. This model states that

F_a is proportional to the vertical flux of kinetic energy of saltating grains. The vertical flux of kinetic energy of saltating grains is proportional to the horizontal flux of the saltating (hopping) grains, as expressed by the Shao et al. (1993) model in equation 9-2:

$$F_a = Km_d(g/\psi)qf(V_H/u_*), \tag{9-2}$$

where ψ is binding energy, K is a constant, m_d is mass per particle, $f(V_H/u_*)$ is a nondimensional function and $q = \int_0^\infty CV_H(z)dz$ where V_H is horizonal speed of the sand grains.

Owen's (1964) theoretical analysis of saltation showed that V_H/u_* may be regarded roughly as a constant. The value of g (acceleration of gravity) is also roughly constant for the Earth's surface. The model predicts that F_a/q is a function only of m_d/μ, the ratio of the mass per saltating particle to the binding energy. Mass per saltating particle reflects the size distribution of the saltating material; coarse sand would have higher mass per particle than fine sand. Binding energy of the PM_{10} particles to larger grains or in aggregates could vary with the following: differences in the texture of the soil or sediment from which they are eroded (i.e., the particle size distribution of the source), chemical composition, clay mineralogy, salt, organic matter content, and a variety of physical properties of the source material, including the (changing) size distribution of soil aggregates as affected by wetting, drying, freezing, thawing, and erosive processes such as sandblasting. Gillette et al. (1997a, b) presented data in figure 9-6 that shows F_a/q_{tot} results for the measurements made in Texas and California.

F_a/q_{tot} for sandy soils does not appear to be a function of friction velocity, but we do not have enough data for other soil textures to evaluate a relationship with u_*.

Clay soils seem to produce less fine particle flux for a given total mass flux, whereas loamy

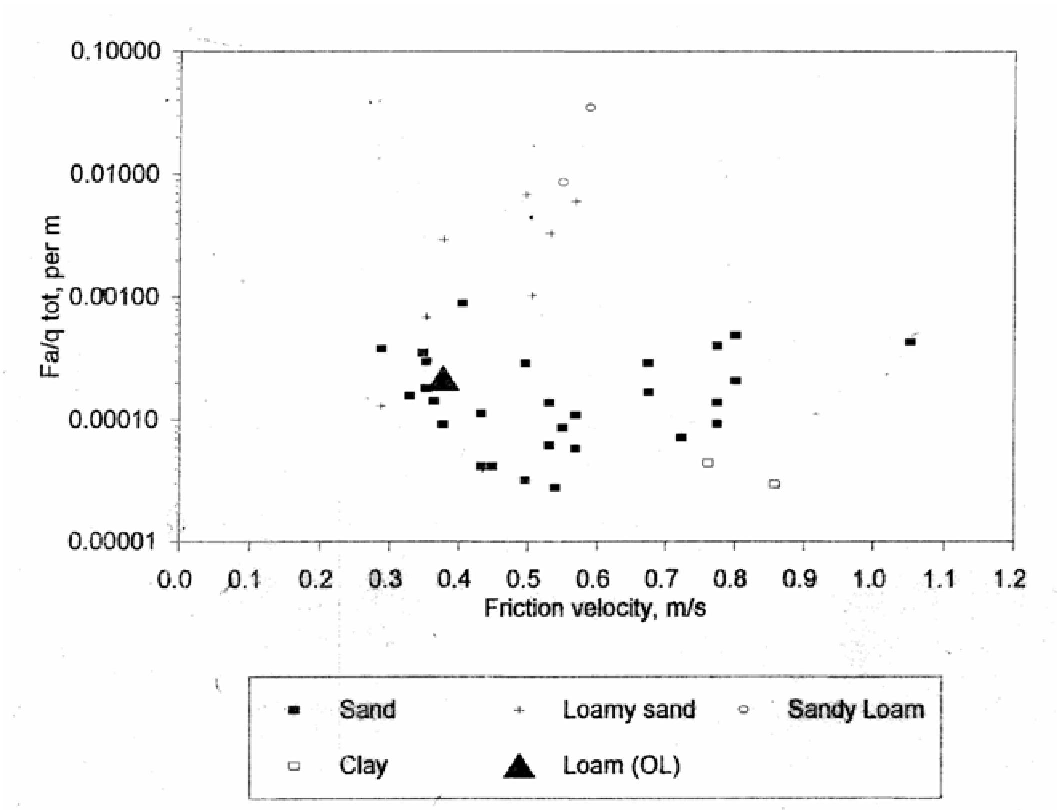

Fig. 9-6. Ratio of vertical flux of particles smaller than 10 µm (PM10) to horizontal flux of total particle mass (q) versus friction velocity for experiments for which different soil textures were found (from Gillette and Chen 2001).

sand soils produce more; the datum from the loam soil is in the middle of the data for sandy soils. The size distributions for q_{tot} values used in figure 9-6 (Gillette and Chen 1999) have a modal value for saltating particles in a range of 120 ± 20 m for all the soils except the clay. The mode of the clay size distribution was roughly three times that for the other soils, which corresponds to a mass about nine times larger. The large modal value for clay corresponds to aggregated particles, not the size of individual clay platelets. Thus μ values are highest for clay, intermediate for sand and loam, and lowest for loamy sand.

Effect of Vegetation on Wind Erosion at the Jornada

Airborne particle fluxes for typical vegetated areas of the Jornada LTER were measured at 15 vegetated sites and at 3 vegetation-free sites on sandy soil. The effect of vegetation was demonstrated by calculating the ratio of the fluxes for the vegetated sites to the mean flux for the three vegetation-free sites. Measurements of particle fluxes at the 15 vegetated sites began in January 1998. These are the sites used for measurements of aboveground net primary production (ANPP; chapter 11)—five vegetation categories having three sites each. The vegetation categories are black grama (*Bouteloua eriopoda*) grassland, creosotebush (*Larrea tridentata*), tarbush (*Flourensia cernua*), playa grassland (characterized by *Pleuraphis mutica*), and mesquite (*Prosopis glandulosa*). Results for January 30 through April 30, 1998 (a period of active wind erosion), are summarized in table 9-2.

Table 9-2. Ratios of the mean particle movements at the vegetated sites to that for the vegetation-free site for January 30-April 30, 1998. Three different plots for each vegetation type were used.

Vegetation Type	Mean Ratio	Standard Deviation
Tarbush	0.0006	0.0002
Black grama grassland	0.0007	0.0003
Playa grassland	0.002	0.002
Creosote	0.002	0.001
Mesquite	0.09	0.06

All three of the mesquite sites were strong sources during this period. Of the five vegetation groups at the Jornada, only the mesquite group is a large-scale dust emitter. The mesquite sites contributed dust at a rate roughly a tenth as large as that from a disturbed bare soil site. Although the other four groups can emit dust during circumstances such as prolonged drought, heavy disturbance, or energetic dust devil activity, they would not normally be significant contributors compared to disturbed bare soil and mesquite sites.

Vegetation acts as a noneroding roughness element that affects the erosion threshold in two ways: (1) directly covering part of the surface and thus protecting it, and (2) absorbing part of the wind momentum that is not then available to initiate particle motion. This momentum partitioning leads to a decrease of the wind shear stress acting on the erodible surface and consequently reduces the erosion efficiency.

Gillette and Stockton (1989) used a scheme developed by Marshall (1971) to calculate the effect of momentum partitioning on threshold friction velocity. Soils having the lowest u_{*t} were the soils most sensitive to changes of vegetation cover. Ratios of threshold (bare)/threshold (vegetated) versus a measure of the geometry of vegetative cover, L_c, were given by Gillette and Stockton (1989). A formula summarizing these data is

$$u_{*t} = \mathrm{u}_{*t(\mathrm{bare})}e^{kLc/2}, \qquad (9\text{-}3)$$

where k is a constant (~ 14). The measure of vegetative cover L_c is the ratio of the frontal silhouette area of plants, as seen by the wind, divided by the total land area, including theplants and bare soil. The experimental data represented by equation 9-3 is for solid hemispheres that were randomly placed in the erodible material. Because the lowest threshold velocity for smooth sandy soils is roughly 20 cm/s and friction velocities larger than 100 cm/s are very rare, a value for L_c of 0.23 should provide good protection for the soil from wind erosion. Assuming a hemispherical model for plant geometry, the fraction of the ground covered by plants (looking down onto the ground) is $2L_c$. Calling this ratio A_c, this model would predict good protection for A_c larger than 0.46. The black grama grassland and playa sites had A_c values of the order 1, so excellent wind erosion protection should be expected. Creosote bush sites do not show significant fluxes in table 9-2 and were visually estimated to have L_c values of about 0.25. Values of this order would be expected to give good protection even on the sandy soils.

All the mesquite sites possessed sandy soils that correspond to vegetation-free threshold friction velocities of 20–30 cm/s. Unlike the other ANPP sites having sandy soil, however, the mesquite sites have significant erosion fluxes. The mesquite sites had A_c values of 0.44, 0.30, and 0.4. A_c values equal to or larger than 0.4 would normally be associated with moderate or better wind erosion protection. The observation of wind erosion where little would be expected according to equation 9-3 led us to suspect that mesquite is a class of vegetation that is not well described by equation 9-3.

One assumption in equation 9-3 is that individual shrubs are randomly oriented in the field. Visual inspection of mesquite at the Jornada suggests a nonrandom orientation of the mesquite plants superimposed on a random orientation; such that a preferred orientation along the direction of strongest wind erosion (southwest–northeast) could exist. Strong areas of wind erosion along with strong areas of deposition were observed along with the orientation.

Marticorena and Bergametti (1995) advanced a physical formulation of the drag partition between the roughness elements and the erodible surface from an approach developed by Arya (1975). The overall threshold friction velocity u_{*t} is expressed as the product of the threshold friction velocity for a smooth surface u_{*ts} (the air–particle interaction of a particle of diameter D_p for a smooth surface having aerodynamic roughness length z_{0s}) and an efficiency factor f_{eff} that expresses the fraction of the wind stress that is available to act on the particle for a surface having aerodynamic roughness z_0 that is rougher than the smooth surface z_{0s} (Marticorena et al. 1997). Marticorena and Bergametti (1995) showed that

$$u_{*ts}/u_{*t} = 1 - (\ln[z_0/z_{0s}]/\ln[0.35(10/z_{0s})0.8]). \qquad (9\text{-}4)$$

Thus, a rough surface increases the overall threshold friction velocity above that for a smooth surface that has the same particle size distribution. The relationship of u_{*t} versus z_0 was confirmed by data of Marticorena et al. (1997). For a sandy soil having a smooth threshold friction velocity of about 20 cm/s, roughness lengths larger than about 0.2 cm would result in overall friction velocities of more than 150 cm/s. Consequently, surface roughness leading to roughness lengths larger than 0.2 cm provides good protection against wind erosion.

The tarbush sites have measured z_0 values that were larger than 0.2 cm. In addition, the bare silty soils between plants had threshold velocities larger than 100 cm/s. In some cases, the soils between tarbush plants had CLCs. For the soils with CLCs, threshold friction velocities were largely beyond the power of our equipment. The above-surface conditions combined make tarbush areas at the Jornada unlikely source areas for dust. Equation 9-4 was not applied to the creosotebush, black grama grassland, or playa sites because we had no measurements of z_0 at those sites. Values of z_0 could be roughly estimated for mesquite sites because for areas of abundant roughness, a linear relationship exists between z_0 and height of the roughness elements (bushes). However, application of equation 9-4 to the mesquite data of table 9-2 also leads to poor predictions. Values of z_0 for the mesquite sites are estimated to be in excess of 1 cm. Strong protection predicted by equation 9-4 for such heights was not observed for mesquite.

For application of equation 9-4, it seems that there needs to be a clear distinction of local z_0 and the z_0 appropriate for an ensemble of vegetation and smooth soil. In mesquite areas, there are local areas of several meters wide by tens of meters long that may have very small z_0 values. However, when taken as a whole, the area containing mesquite shrubs may have z_0 well in excess of 1 cm. When wind conditions are right, local areas with small z_0 values are activated in the mesquite dunes even though the area as a whole would be predicted by equation 9-4 to not have wind erosion.

In short, the Jornada possesses a kind of vegetation (mesquite-dominated) that is a relatively poor protector of the soil. Use of existing theory on the protective effect of mesquite does not predict the rather significant amounts of wind erosion observed at the three Jornada Mesquite aboveground net primary productivity (ANPP) sites. For the other kinds of vegetation at the Jornada ANPP sites (see chapter 11), however, the theory seems to agree with observations that protection of the soils is adequate.

Is the Jornada a Source or a Sink of Eolian Materials? What Materials Are Lost If It Is a Source of Eolian Materials?

From the above work on vegetation versus wind erosion in each major type of vegetation, it seems that mesquite areas and disturbedut are more likely to be depositional areas for dust. To assess whether the Jornada is a net source areas are significant source areas of dust. Areas having other kinds of vegetation might occasionally be source areas b or sink of eolian materials, we need to know the proportion of land covered by mesquite. We must also know the large-scale atmospheric deposition at the Jornada (i.e., how much dust is deposited at the Jornada that comes from large distances). To make this estimate, we need a large-scale emissions/transport/deposition model. Until the model is ready, however, estimates of the vertical flux of eolian material that is carried beyond the borders of the Jornada Basin will be deduced from historical data.

Historical Deposition Rates of Eolian Material

In 1962, seven dust traps consisting of 30- by 30- by 5-cm pans filled flush with 1-cm diameter glass marbles were placed at a height of 90 cm (Gile and Grossman 1979). They were placed in the field each year for 11 years from February through June (the dusty season of the year). The traps were sampling in the dustiest time of the year, and the catch for that interval was considered the total year deposition. Results of deposition averaged for 11 years (1962–72) are shown in table 9-3 along with the size distribution of the particulate material. Using an average bulk density of 1.4 g/cm^3 soil would accumulate at a rate of 2.4 cm per 1,000 years. Thirty-four percent of the deposited material is silt-sized and 24% is clay-sized material.

Table 9-3. Mass deposition in the Jornada Basin, 1962--1972 (after Gile and Grossman 1979).

Trap No.	Deposition g/m^2 /yr	Particle size distribution in % by mass (mm)				
		2-.25	0.25-0.1	0.1-0.05	0.05-0.002	< 0.002
4	15.7	2	14	14	43	27
5	26.3	2	20	22	34	22
6	58.6	4	27	21	26	22
Average	33.5	2.7	20.3	19	34.3	23.7

Historical Mass Vertical Flux of Eolian Materials

Gibbens et al. (1983) estimated the long-term gross erosion rates and net soil loss rates for three sites at the JER. For one of these sites, Hennessy et al. (1986) determined the size distribution for the soil material that had been lost to the site. This site was labeled the natural revegetation exclosure. Mean rates of net soil loss for 1933–80 were established by measuring the change of level on grid and transect stakes. The mean rate of soil loss per year for the deflated areas at the site was 5,200 g/m^2/yr, whereas the mean rate of net soil loss (for the entire area) for the same site was 1,400 g/m^2/yr. Hennessy et al. (1986) determined that there was almost no net loss of sand from the site as a whole; silts (84%) and clays (16%) accounted for all the soil lost. The mean rate of deposition for particles smaller than 50 μm (silt and clay) for the three sites cited in table 9-3 is 19.43 g/m^2/yr. This rate of deposition is much smaller than the mean source rate at the natural revegetation exclosure site of 1,400 g/m^2/yr. Because the area of sandy soil covered by mesquite vegetation is a significant fraction of the JER (more than 10%) and the rate of emission is 73 times the rate of deposition, the Jornada is almost certainly a source area for dust.

For prolific sources of soil dust like the centers of farm fields during dust storms, parts of the Sahara Desert, and parts of Owens (dry) Lake, California, the flux is limited by the momentum from the wind. Gillette and Chen (2000) have found that the supply-limited, vegetation-free source at the Jornada is approximately one-third as emissive as supply unlimited sources. The mesquite-covered, sandy soils of the Jornada would therefore be expected to have emissions rates of the order of 3% (one-third of 9%) of the above prolific dust producers.

Gillette et al. (1974) found that long-distance transport of emitted soil particles occurs for particles whose sedimentation velocity is less than one-tenth of the vertical root mean square velocity. Practically speaking, this corresponds to particles smaller than about 10 μm. Patterson and Gillette (1977) showed that the typical size distribution of wind erosion particles smaller than 10 μm is roughly log normal with a number mode at 0.48 μm and geometric standard deviation of 2.2. Gillette et al. (1978) showed that aircraft-obtained size distributions of dust in southeastern New Mexico dust were quite similar to what Patterson and Gillette (1977) described as typical size distribution for dust storms. Pinnick et al. (1985) determined size distributions for blowing dust at White Sands Missile Range (less than 100 km from the Jornada). They also concluded that the dust distributions were similar to what Patterson and Gillette (1977) described as bimodal log normal distributions having “about the same mode radii.” Pinnick et al. (1993) gave size distributions parameters for the log normal distribution of particles between 0.2 and 30 μm for a dust storm at Orogrande, New Mexico (also less than 100 km east from the

Jornada), and those parameters (0.7 µm and 2.2) were very similar to observations by Patterson and Gillette (1977).

Erosion Rates Measured on Vegetation-Free, Sandy Soil Locations

Figure 9-7 shows the abrasion of the crusted surface at the instrumented and cleared site at three measuring locations spaced at 30-m intervals.

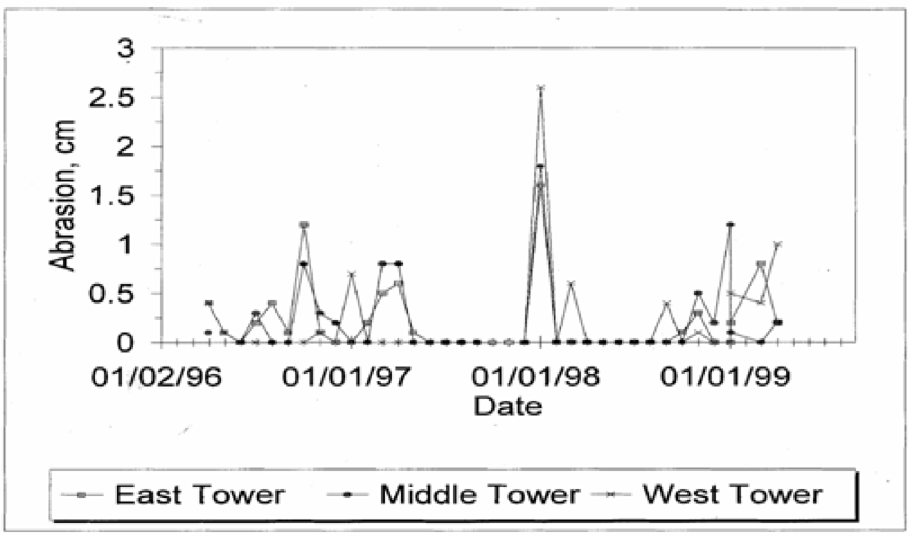

Fig. 9-7. One-month abrasion of the surface crust (progressive distance of the crust surface from an unchanging height above ground) for the three towers at the vegetation-cleared site (from Gillette and Chen 2001).

Figure 9-8 shows the mass fluxes at the same three locations at different points in time. Note in January 1996 an erosion event removed 0.8, 1.7, and 2.5 cm of material from the surface. These large values compare to a mean annual lowering of the soil surface by 1.97, 2.9, and 3.3 cm/yr. Using an average bulk density of 1.6 g/cm^3 measured from seven crust samples obtained at the location, the annual mean masses of surface material lost may be calculated. Multiplying these annual

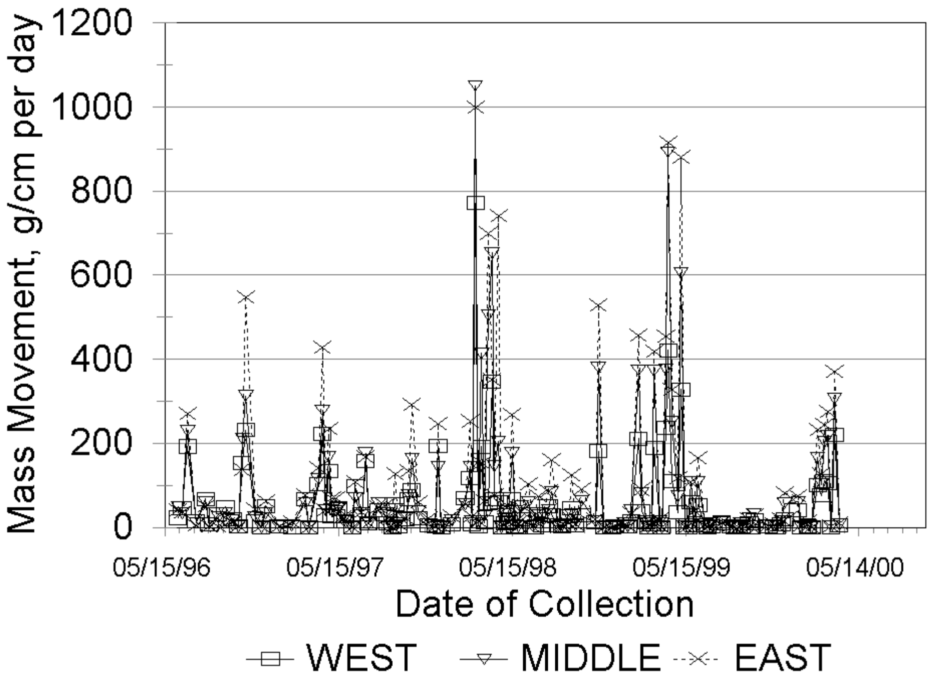

Fig. 9-8. Total particle mass movement for the three towers at the vegetation-cleared site (airborne particles moving parallel with the wind and perpendicular to the ground) in units of g/cm (from Gillette and Chen 2001).

mean mass losses by the ratios of silt and clay to total soil mass for the west, middle, and east sites (0.197, 0.149, and 0.118, respectively) gives the annual mean vertical loss of silt and clay particles. These loss estimates are 6,140, 6,904, and, 6,106 g/m^2/yr.

These emissions are larger than the annual mean net emission rate of 1,400 g/m^2/yr reported by Gibbens et al. (1983) above; however, the instrumented site is bare, and the sites evaluated by Gibbens and Beck (1988) were protected by mesquite plants. The ratio of the long-term loss of silt and clay reported by Gibbens et al. (1983) for the mesquite site to the 2.3-year average for this bare site is 0.22. This ratio is more than twice the mean ratio of 0.09 reported in table 9-2 (the ratio of erosion at the mesquite sites to erosion at the vegetation-free sites), but we consider this to be fair agreement.

From visual inspection of large areas of sand deposition near the vegetation-free site, much of the eroded sand was deposited downwind within about 300 m of the point of emission. The loss of silt and clay from the vegetation-free site is about 5–10 times higher than the loss of silt and clay from the mesquite areas. However, the area of disturbed land similar to the vegetation-free site (for example roads, cattle containment areas, and other disturbed areas) is small, so that the important source of windborne dust is the sandy soil covered by mesquite.

How Does Wind Erosion Change the Soil-Forming Process?

Wind is a major geomorphic force in the Chihuahuan Desert and is responsible for many of the soil patterns in sandy areas (Gile 1966a). The Jornada region contains soils that when traced laterally can be seen in their eroded state, their partially eroded state, and their uneroded state (figure 9-9a). Continued tracing of these soils into areas where eolian sediments have accumulated, such as coppice dunes, shows the effects of progressive burial (figure 9-9b).

As described by Simonson (1959), there are four soil-forming processes that transform material at the Earth’s surface. These are losses, additions, translocations, and transformations. In erosional settings (figure 9-9a) losses are dominant and responsible for the truncation of many soil profiles in the Chihuahuan Desert (Monger 1995). Because the A horizon (i.e., topsoil) of any soil contains much of the humified organic matter, N, available P, microbial population, and the seed bank, the erosion of this layer results in a loss and redistribution of these biotic components (e.g., Schlesinger et al.

1990). Furthermore, because water-holding capacity is largely controlled by organic matter and silt content (Herbel et al. 1994), removal of these constituents will decrease a

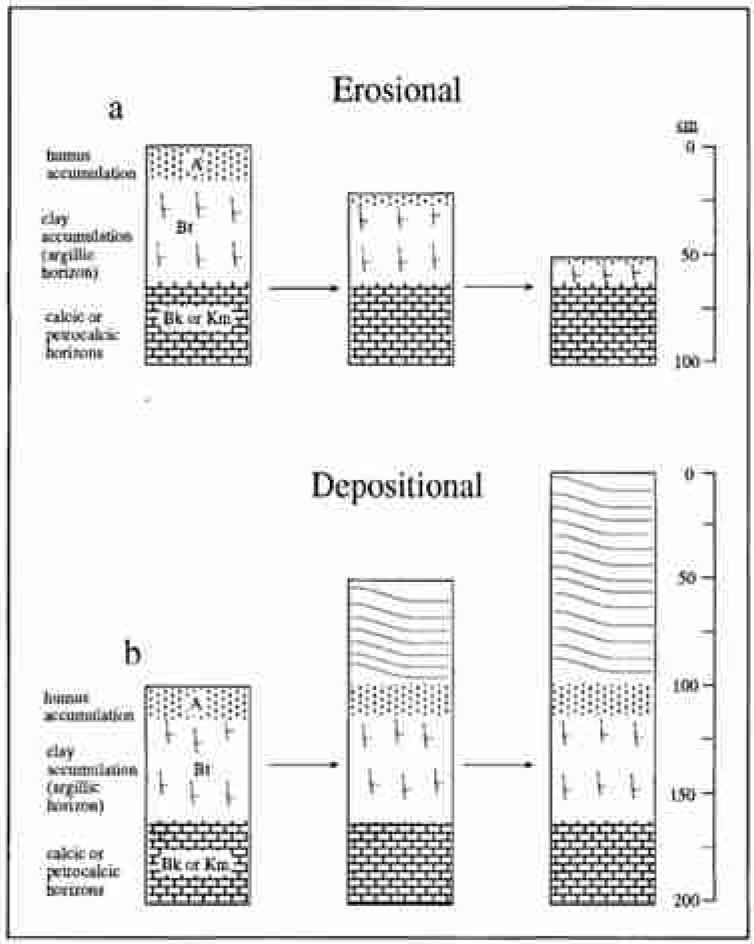

Fig. 9-8. Total particle mass movement for the three towers at the vegetation-cleared site (airborne particles moving parallel with the wind and perpendicular to the ground) in units of g/cm (from Gillette and Chen 2001).

soil's ability to retain water. Organic matter is also important for aggregate formation (Brady and Weil 1996). Its decline therefore leads to a decline in infiltration, which in turn leads to greater runoff (Bull 1991), causing hill slope erosion in addition to wind erosion and the perpetuation of bare ground. This bare ground is not only susceptible to further erosion but also a thermally harsh environment in which seedlings do not easily become established (Davenport et al. 1998).

For soils with calcic and petrocalcic horizons, truncation of the profile leaves those horizons near or at the surface. When this happens, the horizons degrade physically as the result of root growth and burrowing animals (Gile 1975b) and degrade chemically by dissolution by percolating waters. Theoretically, this situation would cause the calcic or petrocalcic horizons to change from a reservoir of atmospheric CO_2 to a source of atmospheric CO_2 unless the dissolved products make it to the groundwater or the ocean. Another important aspect of eroded calcic and petrocalcic horizons is their role in water storage. Microporosity in calcic and petrocalcic horizons causes water to be held more tenaciously than water is held in soil material with larger pores (Hennessy et al. 1983b). Consequently, calcic and petrocalcic horizons are important for preserving sources of water below the layers during droughts for both grasses (Herbel et al. 1972) and shrubs (Cunningham and Burk 1973).

Also in the erosional settings illustrated in figure 9-9a, the other three soil-forming processes are very much modified by the losses process. Because of increased bare ground, additions to the soil profile, in the form of dry dust, wet dust, ions in rain, and organic matter, would not be retained as readily as on noneroded soils covered with vegetation. Because of greater runoff, translocations of particles and ions into and through the soils are curtailed in bare soils (Gile et al. 1969). For the same reason, transformations, such as chemical weathering and carbonate formation, would diminish in comparison with vegetated noneroded soils.

In the depositional settings (figure 9-9b), the four soil-forming process are dominated by the additions process. In this case, aggradation of the land surface lifts the depth of wetting progressively higher in the soil profile. Consequently, the zone of carbonate formation, the zone of clay accumulation, and the zone of organic matter accumulation rise with additions of accreting sediments. Based on the coarse texture of coppice dunes (Gile et al. 1981), most of the additions are in the form of saltating sand particles. However, suspended particles, organic matter, and associated nutrients also accumulate, as well as ions in rain funneled into the soil along plant stems (Whitford et al. 1997). Cumulatively, these additions produce one form of the islands of fertility described by Schlesinger et al. (1990; 1996).

Losses in the depositional setting are mainly derived from silt winnowed from sand (Hennessy et al. 1986). Little evidence of translocations has been found in coppice dunes (Gile 1966b; Gile and Grossman 1979). The absence of translocated material, however, reflects the young age of the deposits (Buffington and Herbel 1965). In some cases, 86.9 cm of sandy sediments have accumulated in 45 years (Gibbens et al. 1983). Similarly, transformations are minor in coppice dunes because of their young age. However, some biomineralization of calcite on mesquite roots and associated fungal hyphae has occurred, as well as decomposition of plant litter.

Conclusions

Based on the physical principles that the wind erosion model incorporates, we can give provisional answers to the questions asked in the first section.

1. From our measurements at the JER, undisturbed sandy soils populated by mesquite and disturbed soils of all types would be expected to be erodible. Other soils are erodible, but only at very high winds that would be experienced only rarely. The most vulnerable soils are the sandy soils that make up about half of all the Jornada Basin soils. The silt soils, playas, and gravelly soils are less vulnerable to wind erosion, and clay soils are less vulnerable than sand soils but more vulnerable than silt/playa/gravel. 2. Wind erosion at the Jornada is governed by several physical mechanisms. Beyond a threshold wind speed, mass flux increases at about the cube of the wind speed. Other important variables affecting wind erosion are length of bare soil between vegetation, soil crusting (biological and physical) and soil moisture, which has a small but measurable effect in reducing dust flux.

3. Change of vegetation cover leads to a change of the source/sink relationship. Grass is one of the most effective protectors of the soil with respect to wind erosion. When grass is replaced by mesquite plants surrounded by large areas of bare soil, wind erosion increases dramatically. Wind erosion also increases dramatically when grass cover is reduced by drought or disturbances. Vegetation patterns are very important determinants of wind erosion on all soil types.

4. It is highly likely that the Jornada Basin is a net source of eolian materials. Historical studies show that soil loss rates for mesquite areas having sandy soils (1,400 $g/m^2/yr$) are much greater than the average dust deposition of 19.4 $g/m^2/yr$. Historical soil losses are roughly consistent with current soil loss rates.

5. Wind erosion changes the soil-forming process in both erosional and depositional settings. In erosional settings, soil profiles are truncated, organic matter and

nutrients are removed and redistributed, infiltration decreases, water erosion increases, and bare ground is perpetuated by physically harsh conditions in intershrub spaces. In depositional settings, sandy sediments from which silts have been winnowed accumulate and progressively bury existing soil profiles. Consequently, the zones of carbonate accumulation, clay accumulation, and organic matter accumulation are lifted with the accreting sediments. Soil resources are more readily trapped and retained in depositional settings.