Article
Contr. Atmos. Phys., February 1999, p. 67–77
0005-8173/99/01 0067-11 $ 3.00/0
Vol. 72, No. 1

A Qualitative Geophysical Explanation for "Hot Spot" Dust Emitting Source Regions [1]

DALE A. GILLETTE

Atmospheric Sciences Modeling Division, Air Resources Laboratory,
National Oceanic and Atmospheric Administration, Research Triangle Park, NC 27711, USA

(Manuscript received March 11, 1998; accepted June 03, 1998)

Abstract

A survey of the sources of dust given by Professor J. Prospero in the Alfred Wegener Conference showed that rather than homogeneous large areas of dust emissions, much of the global supply of dust comes from relatively small (compared to the size of deserts) consistently active dust producing areas. These areas were named "hot spots". To explain these hot spots, I have reviewed what is known about the production mechanisms of dust. From this review I would offer that the primary geophysical explanation of "hot spots" is lack of nonerodible roughness elements, low threshold friction velocity, and lack of aggregation or crusting. These conditions often apply in depositional environments.

1 Introduction

Previous field studies and remote sensing studies have pointed out that the sources of dust carried globally significant distances are not homogeneous over large areas. J. Prospero, in his presentation at the Alfred Wegener Conference compared the results of in situ measurements of aerosols with remotely-sensed aerosols properties obtained from the AVHRR over oceans (Husar et al., 1997) and TOMS (Herman et al., 1997) satellite instruments of absorbing aerosols above 2 km. An important conclusion from these studies is that mineral dust is a substantial (and often major) aerosol species over large ocean regions (Husar et al., 1997). The TOMS absorbing aerosol satellite product shows that much of the dust appears to derive from well-defined sources that are persistently active, year after year (Herman et al., 1997). These "hot spots" in the TOMS aerosol product are not necessarily found in hyper arid regions; rather, the most productive sources appear to be modern and ancient depositional environments (e.g., lake and river beds; deposits in regions of runoff from elevated terrains) located in transitional arid regions. These "hot spots" are often part of "source regions" that for a large extent are "hot spots" surrounded by areas of much lower dust production. On a smaller scale, aerial photographs of agricultural lands in West Texas USA show that a very small fraction of the fields actually produces visible dust plumes (see Figure 1). The fields where I studied dust emissions in West Texas (Gillette, 1981) were hot spots: intense areas of dust production surrounded fields where little if any dust was being emitted. It is the purpose of this paper to summarize mechanisms of dust production that may explain the existence of these "hot spots" (localized intense dust production areas that persist in time) within a larger source area. Some mechanisms have already been elucidated in other publications. Indeed, Pye (1987) has summarized several mechanisms that would explain the formation of concentrated areas of increased wind erosion. For example, salt weathering, chemical weathering, and frost weathering all contribute smaller and more erodible particles in certain environments. My intent is to qualitatively explain dust production "hot spots" from substantiated data rather than to hypothesize new mechanisms.

Table 1 gives a listing of mechanisms leading to wind erosion and their associated effects. The scales of these mechanisms range from synoptic, meso to the micro. The first grouping of mechanisms is Meteorological (I).

[1] A contribution to Alfred-Wegener-Conference "Sediment and Aerosol" in Leipzig, Germany, March 10–12, 1997

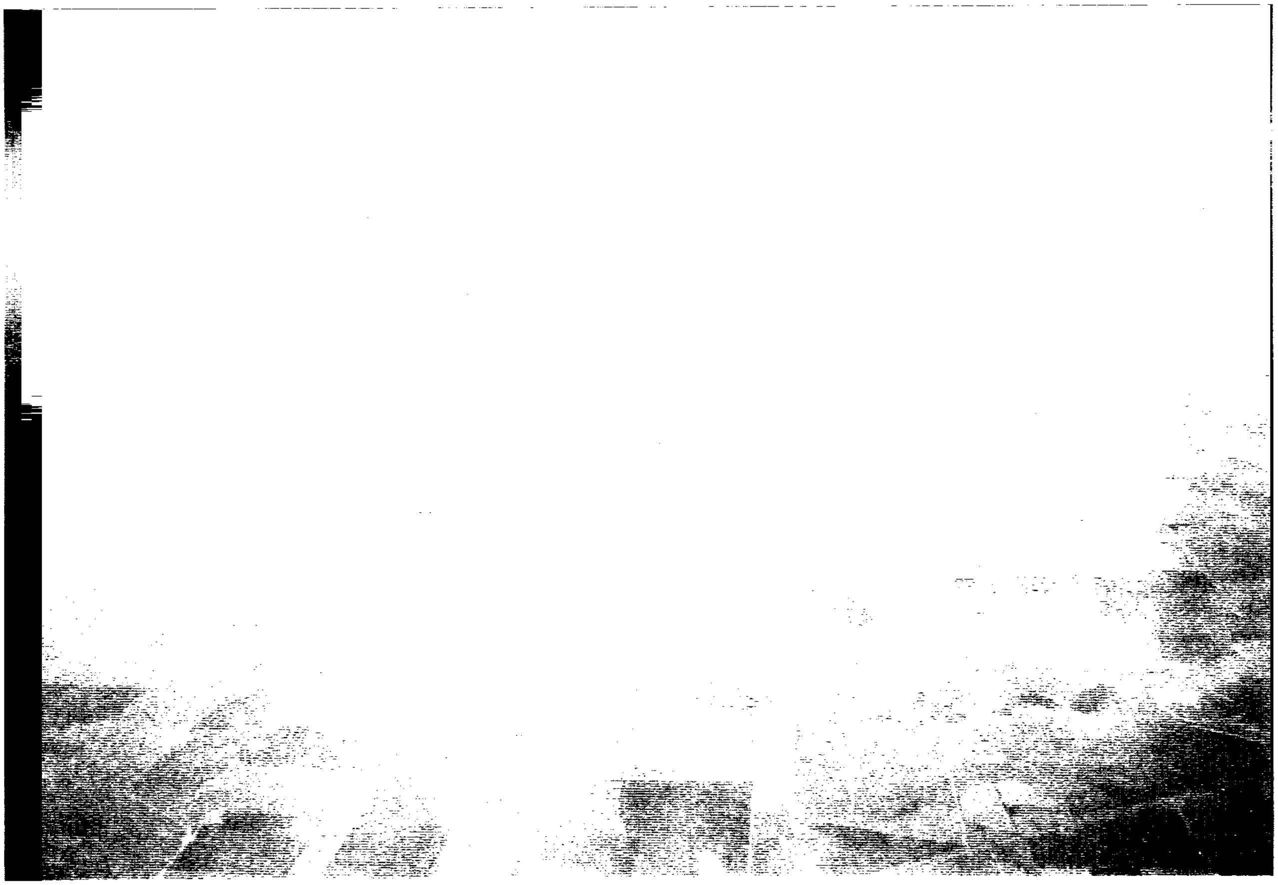

Figure 1: Aerial photograph taken in 1975 during a dust storm in West Texas, USA. Notice that visible plumes are formed over a small fraction of the total amount of agricultural land.

Table 1: Mechanisms of Dust Production by Wind Erosion.

Mechanism	Effects
I. Meteorological	Surface winds: momentum source of erosion, regional high winds
A. Synoptic scale	
B. Meso-scale	Modification of wind by topography, thermally driven circulations
C. Local-scale	Small-scale wind erosion, e.g., dust devils (not persistent) or small scale topography (persistent)
II. Air-particle interactions	
A. Wind speed pdf and threshold wind speed	
B. Airborne particle effect on flow (Owen Effect)	Aerodynamic roughness height (z_0) changed by airborne particles
III. Interactions of surface/air	
A. Threshold friction velocity	
1. Size distribution of particles	Threshold friction velocity u_{*t} on a smooth surface is minimum (about 22 cm s^{-1} for 60–120 μ m particles
2. Momentum partitioning	Increase of u_{*t} with z_0 effect of clasts, vegetation and microtopography
B. Drag coefficient	Drag coefficient increases with z_0
IV. Changeable physical properties of surface/particles	u_{*t} , z_0 changed; mass flux changed
A. Aggregation-Disaggregation	
B. Crustings/Crust destruction	
C. Trapping	
D. Particle supply limitation	
E. Soil moisture	

2 Meteorological Mechanisms

Synoptic scale and meso-scale meteorological systems deliver momentum to the surface in a variety of forms. An example of a synoptic-scale structure that is often associated with wind erosion is the prefrontal wind storm. Large-scale systems do not explain the existence of local "hot spots" since strong dust production is not uniformly observed for the entire land surface over which the system passes. Meso-scale structures such as haboobs (downdrafts of thunderstorms) create short-lived intense local dust production, but are short lived, and may cause erosion in locations that do not normally produce dust; that is our definition of a "hot spot" includes persistence in time for production of dust from a given location. A meso-scale thermal circulation that leads to a "hot spot" was described by Tapper (1991) as circulation that develops over dry salt lakes. However, in order for a "hot spot" to be produced, the strong local winds produced by the circulation must be located over erodible soils. Hot spots may also be caused by topography influencing synoptic-scale circulation. An example of topography-influenced high winds is the funneling of northerly and southerly winds at Owens Lake, California. At this dry lake, large areas of erodible soil underlie the converging air caused by channeling by the Sierra and Inyo mountains. Small scale circulations, for example dust devils and thermally driven local circulations caused by differences in albedo or local topographic features cause local wind erosion. For dust devils, the locations of the dust devil funnels on the land surface are almost random and shifting in time, thus do not cause a "hot spot" at a specific location. As in all the other meteorological mechanisms providing strong winds at the surface, the surface must possess erodible soils for there to be a "hot spot" In the present paper, we consider meteorological mechanisms that produce strong winds to be necessary but not sufficient to explain a "hot spot".

3 Air-Particle Interactions

3.1 Influence of the Wind Speed Probability Distribution and Threshold Friction Velocity

The influence of the wind probability distribution is shown in Eq. (3.1). In Eq. (3.1) E is the expected saltation flux (particles moving in hopping motions), U_t is the threshold velocity, erosion increases as the cube of wind speed U, and $f(U)$ is a probability density function of the wind speed. Here c_D is the drag coefficient, k' is a constant, ΔT is the time interval considered, and L is the length of the source area measured parallel the wind direction. Threshold velocity is that wind velocity at which the erosion process starts; for wind speeds lower than threshold, there is no erosion. It occurs when aerodynamic lift and drag equal forces of gravity and cohesion holding particles in position. Because the wind erosion proceeds as the cube of wind speed after threshold velocity is surpassed, the probability distribution of wind speed is very important.

$$E = k' c_D^{1.5} \frac{\Delta T^{\infty}}{L} \int_{U_t} U^3 f(U) dU \tag{3.1}$$

To illustrate Eq. (3.1) using a simple but plausible probability density function, $f(U)$ is taken to be the Rayleigh distribution given in Eq. (3.2).

$$f(U) = \frac{\pi U}{2\overline{U}^2} \exp\left(\frac{-\pi U^2}{4\overline{U^2}}\right), \tag{3.2}$$

where $\overline{U}$ is the mean wind speed.

The Rayleigh distribution successfully simulates empirical probability density functions of wind speed in many areas having wind erosion problems (Cowherd et al., 1984). A solution for Eq. (3.1) is

$$E = k c_D^{1.5} \Delta T \left(\frac{4}{\pi}\right)^{1.5} \overline{U^3} \left(1 - \Gamma(2.5, z)\right) \tag{3.3}$$

where $\Gamma(2.5, z)$ is an incomplete gamma function and

$$z = \frac{\pi}{4} \left(\frac{U_t}{\overline{U}}\right)^2 \tag{3.4}$$

For a given location, the mean wind speed is a constant so that the variable part of the right side of Eq. (3.3) is proportional to the complement to the incomplete gamma function. Details on the incomplete gamma function are given by Abramowitz and Stegun (1964). This allows us to plot the relative erosion versus the ratio of threshold wind velocity to average wind velocity in Figure 2.

Figure 2 shows that for the ratio of threshold wind velocity to average wind velocity less than 0.5, the total erosion is almost constant. For the ratio of threshold wind velocity to average wind velocity greater than one, however, expected saltation decreases very rapidly.

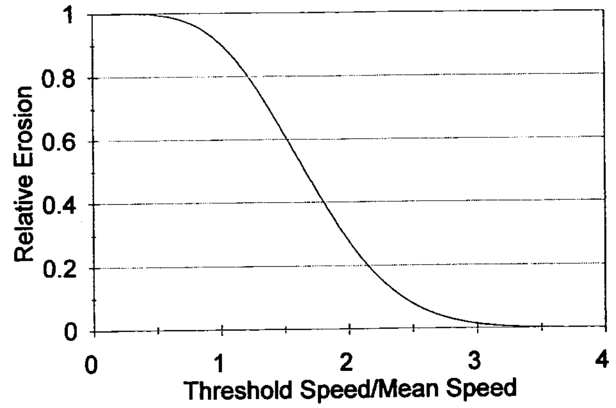

Figure 2: Relative wind erosion from Eq. (3.1) versus the ratio of threshold wind speed to mean wind speed.

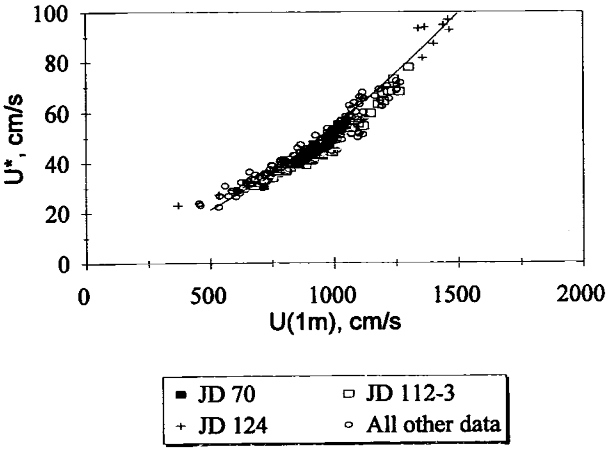

Figure 3: Wind friction velocity $u_\star$ versus wind speed U at 1 m for a location at Owens Lake. The solid curve is a regression of the data. The straight line portion of the curve corresponds to a constant drag coefficient. The curved portion of the curve shows the "Owen Effect" where saltating particles effectively increase the aerodynamic roughness height z_0 of the surface (after Gillette et al., 1997).

The influence of the probability density function (pdf) of wind speed and threshold velocity for saltation flux on the intensity of wind erosion is enormous. For the Rayleigh pdf, Eq. (3.3) shows that saltation flux may be thought of as a potential depending on mean wind speed cubed times a power of the drag coefficient and times a function of mean wind speed and threshold wind speed. This function (given in Eqs. (3.3) and (3.4)) is very sensitive to the threshold wind speed.

3.2 The Owen Effect

To remove the variability of wind speed with height, it is customary to use friction velocity $u_\star$ rather than wind velocity. The relationship $u_\star$ to U for conditions of no erosion, is given by Kaimal and Finnigan (1994) as

$$U(z) = \frac{u_\star}{k} \ln\left(\frac{z}{z_{ONS}}\right) \tag{3.5}$$

where $U(z)$ is wind speed at height z, $u_\star$ is friction velocity, k is von Karman's constant (0.4) and z_{ONS} is aerodynamic roughness length for non-saltating conditions.

A feedback mechanism caused by the hopping motions of saltating particles (the Owen Effect) is a secondary cause of variability. It is a mechanism by which airborne sediment increases the aerodynamic roughness height of the surface. The effect is most easily detected as an increase of the ratio of friction velocity $u_\star$ to mean wind U — see Figure 3. Wind friction velocity $u_\star$ is plotted versus wind speed U at 1 m for a location at Owens Lake. The solid curve is a regression of the data. The portion of the curve for lower wind speeds ($U(1\text{m})$) corresponds to a nearly constant drag coefficient. The portion of the curve for higher wind speeds shows the "Owen Effect" as a curvature of the $u_\star$ versus $U(1\text{m})$ data point loci where saltating particles effectively increase the aerodynamic roughness height z_0 of the surface.

Raupach (1991) incorporated the theory of Owen (1964) to provide an expression for the aerodynamic roughness length, $z_{0\text{salt}}$, when saltation occurred on smooth or rough surfaces. The expression for $z_{0\text{salt}}$ is equal to aerodynamic roughness length (z_{ONS}) for wind speed below the threshold of particle movement and is roughly equal to Owen's aerodynamic roughness length for saltation on smooth surfaces. For conditions when wind erosion is occurring, the aerodynamic roughness length during saltation $z_{0\text{salt}}$ is given by the Raupach formula as:

$$z_{0\text{salt}} = \left(A\frac{u_\star^2}{2g}\right)^{1-R} z_{ONS}^R \tag{3.6}$$

where A is a constant of order 1 and $R = u_{*t}/u_*$ where u_{*t} is the threshold friction velocity for wind erosion and g is the acceleration of gravity.

The Owen Effect causes an amplification of the differences already caused by nonhomogeneous threshold friction velocity. The Owen Effect locally increases the drag coefficient. For a region having a uniform large scale wind, the Owen effect can cause friction velocities to be larger for areas that are eroding than for areas that are not eroding.

In explaining the existence of dust source "hot spots", the Owen Effect dramatically increases the flux with distance. One example of this was documented by Gillette et al. (1997). That is, source areas having long fetches of low-threshold friction velocity particles parallel the wind direction for high winds will produce larger dust fluxes than equal areas where the fetches are smaller.

4 Interactions of the Surface with the Air

4.1 Threshold Friction Velocity

4.1.1 The Influence of Size Distribution of Surface Material on Threshold Friction Velocity

Iversen and White (1982) gave formulae for threshold friction velocity for particles on a smooth surface of a wind tunnel various particle densities (0.21 to 11.35 g cm^{-3}) and diameters (12 to 1290 μm). For particle densities equal to 2.65 g cm^{-3} the formulae of Iversen and White showed a minimum friction velocity for the particle size range corresponding to the size range 60–120 um diameters.

Observations of Gillette et al. (1980), Gillette et al. (1982), Gillette (1988), Chatenet et al. (1996), and Marticorena et al. (1997) were that for most wind erosion source areas investigated except erodible gravel surfaces and clay-textured sediments, particles of the diameter 60–120 μm are found. Since the threshold velocity is that of this size of the available particles found in the surface soil, the smooth surface threshold for erodible soils is about 22 cm s^{-1}. Some clay-textured sediments possess loose particles of the size 60–120 μm diameters (Gillette, 1988). However, most clay sediment surfaces had larger loose particle sizes and therefore required higher friction velocities to start wind erosion. Source areas having loose particles of the size 60–120 μm diameters are optimal for producing dust since they have the lowest threshold velocities. For a given region where wind is roughly constant, Eq. (3.3) shows that wind erosion is largest for the smallest threshold velocity. These areas have been largely identified with sediments having composition mostly sand. Soils having a different composition may be found to have loose particles of the size 60–120 μm diameters, however. Many highly erodible soils of the "dust bowl" in the United States during the 1930's were loamy soils (for example loams and sandy loams) disaggregated by agricultural working (disturbance) during periods of intense and sustained drouth. These soils possessed great quantities of loose particles of size 60–120 μm diameters although during normal times, they were more aggregated. Disturbance of clay soils also produces copious quantities of loose particles of size 60–120 μm diameters. Consequently, hot spots can be located in areas having large quantities of loose particles of size 60–120 μm diameters – whether by sandy soils or by disturbed silty or clayey soils.

4.1.2 Surface Roughness Effect on Aerodynamic Roughness Height: Momentum Partitioning

Surface nonerodible roughness acts to increase threshold friction velocity above values found for smooth surfaces. This is done in two ways: (1) nonerodible roughness elements directly cover part of the surface; (2) they also absorb part of the wind momentum that would have been available to initiate particle motion. This momentum partitioning leads to a decrease of the wind shear stress acting on the erodible surface and consequently the erosion efficiency.

Gillette and Stockton (1989) used a scheme developed by Marshall (1971) to calculate the effect of momentum partitioning on calculating the effect on threshold friction velocity. Musick and Gillette (1990) quantified the momentum partitioning to vegetation by using the same scheme based on Marshall's experiments as was used for quantifying smaller-scale non-erodible elements. An alternate method was given by Marticorena and Bergametti (1995) to calculated the drag partition between the roughness elements and the erodible surface from an approach developed by Arya (1975). They expressed friction velocity u_{*t} in Eq. (4.1) as the 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 height z_{0s}) divided by an efficiency 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

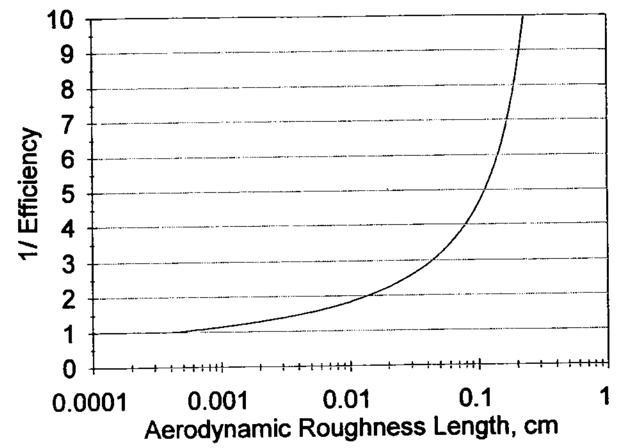

Figure 4: The ratio f_{eff} versus z_0 from Eq. (4.3).

that is rougher than the smooth surface z_{0s} (Marticorena et al., 1997).

$$u_{*t}(D_p, z_0, z_{0s}) = \frac{u_{*ts}(D_p)}{f_{\text{eff}}(z_0, z_{0s})} \tag{4.1}$$

Marticorena and Bergametti (1995) developed the formula

$$f_{\text{eff}} = \frac{u_{*s}}{u_*} = 1 - \frac{\ln\left(\frac{z_0}{z_{0s}}\right)}{\ln\left(0.35\left(\frac{10}{z_{0s}}\right)^{0.8}\right)} \tag{4.2}$$

The value of f_{eff} versus z_0 is plotted in Figure 4. Figure 4 shows that the efficiency for eroding particles is greatly reduced for aerodynamic roughness heights of about 0.1 cm. Thus a rough surface can have a large effect in increasing the overall threshold friction velocity for a surface that has the same particle size distribution as for a smoother surface. The relationship of u_{*t} versus z_0 was confirmed by data of Marticorena et al. (1997). A "hot spot" would not be expected to have a z_0 value larger than 0.1 cm.

Probably the most effective deterrent to wind erosion is momentum partitioning to vegetation and other large non-erodible material. Bare areas are expected to have more intense dust emissions than are vegetated areas. Vegetation and larger objects like cobbles and gravel on the surface absorb momentum that would otherwise go to particles in the surface sediment. Eq. (4.2) can be used for both solid nonerodible elements like cobbles and for vegetation. Field studies of the role of vegetation in protection against wind erosion were conducted by Wolfe and Nickling (1993) and by Lancaster and Baas (1997).

4.2 Effect on Drag Coefficient

Another important effect of a change of surface roughness is on the drag coefficient. The drag coefficient C_d may be defined as

$$C_d^{0.5} = \frac{u_*}{U(z)} = \frac{k}{\ln\left(\frac{z}{z_0}\right)} \tag{4.3}$$

for near neutral atmospheres (Priestley, 1959). Since the mass flux of saltation is roughly proportional to the cube of friction velocity above threshold friction velocity, it is also roughly proportional to the wind speed cubed times $C_d^{1.5}$. An increase of z_0 thus leads to an increase in drag coefficient and a consequent increase in mass flux rate. The variability of mass flux rate caused by the roughness of the surface was identified in experiments performed at Owens Lake (Gillette et al, 1997).

The effect of increasing roughness decreases the flux in Eq. (4.2) (more protection of particles) and increases it in Eq. (4.3) (more momentum flux). The opposite influences caused by the change of roughness cause the dominating effect to not be obvious. Sensitivity studies by Marticorena (private communication, 1998) showed that for low wind velocities the dominant factor is the increase of the threshold friction velocity that leads to a decrease in the saltation flux. Figure 4 shows that the protection by roughness – the effect of increasing threshold is overpowering when z_0 reaches the value of about 0.1 cm. Overall, the z_0 should be less than 0.1 cm for "hot spots".

5 Changing Physical Properties of Surface Particles

5.1 Aggregation / Disaggregation

A large source of the temporal and spatial variability of threshold friction velocities is the aggregation and disaggregation of soils that can change both u_{*ts} and z_0. Aggregation is a mechanism by which individual particles in the surface sediment become effectively larger particles. Aggregation usually occurs after moistening of the surface followed by drying of the surface. Composition such as efflorescent salts, calcium carbonate and gypsum can promote aggregate formation. 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 surface aggregates depends on the availability of sandblasting, the presence of freeze-thaw cycles, efflorescence of salt, formation of crystals and temperature of the sediment. Destruction of aggregates is especially important for cultivated soils where aggregates are clods formed by cultivation. Chepil (1951) stated that stability of aggregates to water is accomplished by water insoluble cements composed of clay particles and irreversible or slowly reversible inorganic and organic colloids. Chepil and Woodruff (1963) recognized the effect of high calcium carbonate in increasing wind erodibility. Gillette (1988) also described the lower threshold friction velocities for soils rich in Calcium Carbonate. These soils form aggregates that are more weakly cemented than soils having the same texture but without calcium carbonate. The competing effect of tendency to form aggregates for calcium carbonate is more than compensated by the weakness of the aggregates that are formed. The application of water is especially effective in destroying water-unstable aggregates when the water is in the form of intense rainfall.

5.2 Crusting / Destruction of Crust

Observations of crust formation and destruction on Owens Lake, California shows extreme differences in the presence of crusted sediment and loose sediment from year to year and from season to season. Crusts form following a thorough wetting by precipitation. Gillette et al. (1982) observed for natural crusts found in Southwest American desert soils that silt and clay sediments form thicker and stronger crusts than do predominantly sandy sediments. For clay soils, Breuninger et al. (1989) observed that increased organic content correlates with lack of crusting and breakage of the massive clay into smaller and smaller peds that are vulnerable to wind erosion. Any process that acts to break crusts or prevent their formation is conducive for the formation of a dust emission "hot spot". Such processes are intense rainfall and hail, freeze/thaw, and thermal expansion. It must be mentioned here that certain clay crusts are broken by curling and distortion of the original surface. For many of these broken clay crusts, however the broken pieces are large enough and hard enough that they resist erosion (Gillette et al., 1982).

5.3 Trapping

Trapping of particles that are moving in saltation reduces the dust flux by wind erosion. Such trapping can happen on surfaces roughened by vegetation, clasts or by ridged soil. Trapping is especially important for agricultural soils where plowing causes both nonerodible surface clods and furrows. Trapping by plowing has been widely described in agricultural literature as a mitigation method for wind eroding fields. Trapping requires large nonerodible elements that would likely cause z_0 to be larger than 0.1 cm. Therefore, trapping will be incorporated in the "threshold z_0" in this treatment: for $z_0 < 0.1$ cm, trapping of particles should not be a major mechanism for inhibiting erosion.

5.4 Particle Source Limitation

Particle limitation (depletion of the loose "available" erodible material on the surface) has been suggested as a cause of decreasing of the dust flux rate. It is illustrated in Figure 5. At an eroding site, the mass flux at 10 cm, (units of mass per square cm per second), versus $u_\star^3$ is shown. The slope of the points at the beginning of the erosion episode "before 11.1 h" decreased sharply "after 11.1 h". Other observation sites simultaneously did not show the decrease of the flux at 10 cm versus $u_\star^3$ slope; nor did the pattern of the mass flux at 10 cm versus $u_\star^3$ points at the same station for other days show the decrease of slope. The interpretation of these data was as follows: "before 11.1 h" represents the particle flux from a layer of loose material; "after 11.1 h" represents the new lower flux rate reflecting crust material that has little available loose-particle mass. Consistent with this interpretation were observations of the sample site as having a general thin, loose sand cover within which were a few large bare patches of crust. The sharp decrease of the slope of mass flux at 10 cm versus $u_\star^3$ from one constant value to another lower value probably coincided with the moving-off of a loose particle layer, leaving behind the loose-particle-limited crust.

5.5 Soil Moisture

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 (1989) showed that sand grains are held together by the capillary effect of soil moisture. However, observations by the author are that vigorous wind erosion can follow as quickly as 10–30 minutes following a soaking rain storm. The reason for such short time mitigation of wind erosion is that the eroding layer need be only about a millimeter thick; the

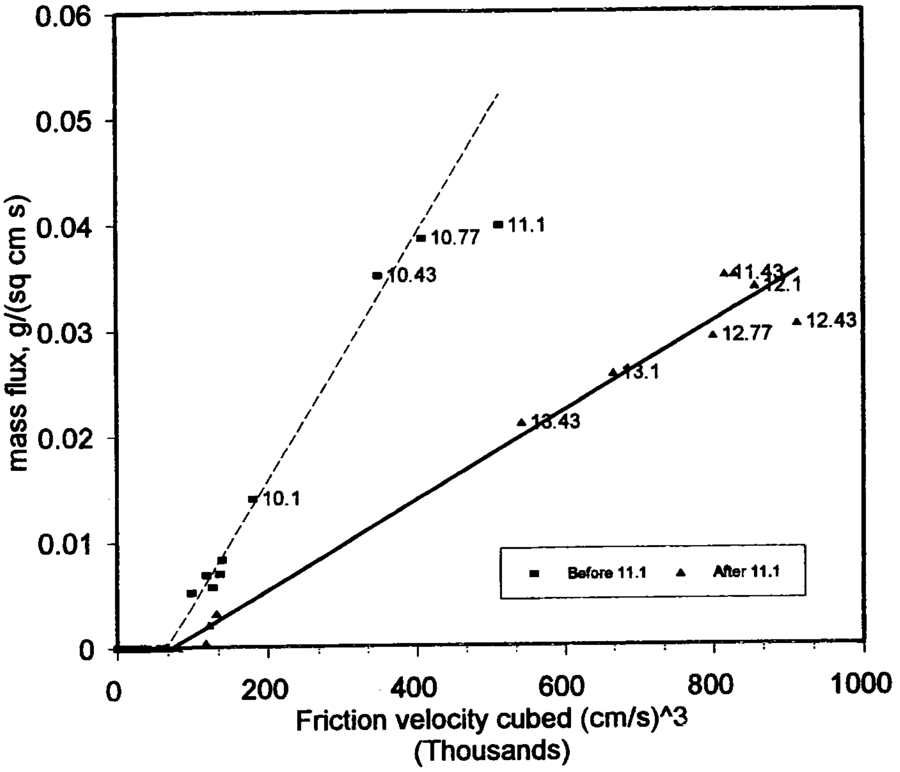

Figure 5: Twenty-minute mass flux at site 5012 versus u_*^3 for Julian Day 124 before and after 11.1 h. Local time is shown for selected data points. The mass flux shows a linear relationship with the cube of friction velocity until 11.1 h, then decreases as u_*^3 increases. After one hour, a new linear relationship of mass flux with u_*^3 is seen but with a smaller response of mass flux for a given u_*^3. We interpret this as a depletion of available particle supply at 11.1 h followed by a smaller but still linear relationship between mass flux and u_*^3 – particle supply limitation. (After Gillette et al., 1997).

time to dry such a thin layer in high wind conditions is quite short. A longer-lasting effect caused by atmospheric precipitation is the formation of crusts and aggregates (see Section 5.1 and 5.2 above), and the growth of vegetation that acts to protect the surface (see Section 4.1 above).

6 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 surface vertical flux of kinetic energy of saltating grains (which causes the release of fine aerosols). 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 Eq. (6.1):

$$F_a = Km_d \frac{g}{\psi} qf\left(\frac{V_H}{u_*}\right), \tag{6.1}$$

where ψ is binding energy, K is a constant, m_d is mass per particle, $f(V_H/u_*)$ is a nondimensional function, where V_H is the horizontal speed of the particles (m s^{-1}) impacting the surface, and

$$q = \int_0^\infty CV(z)dz,$$

where $V(z)$ is the horizontal particle speed (m s^{-1}) at height z, and C is the concentration of saltating (hopping) particles (g m^{-3}). Owen's (1964) theoretical analysis of saltation showed that V_H/u_* may be regarded as roughly a constant. The value of g (acceleration of gravity) is roughly constant for the earth's surface. The above model would then predict that F_a/q is a function only of m_d/ψ, the ratio the mass per saltating particle to the binding energy.

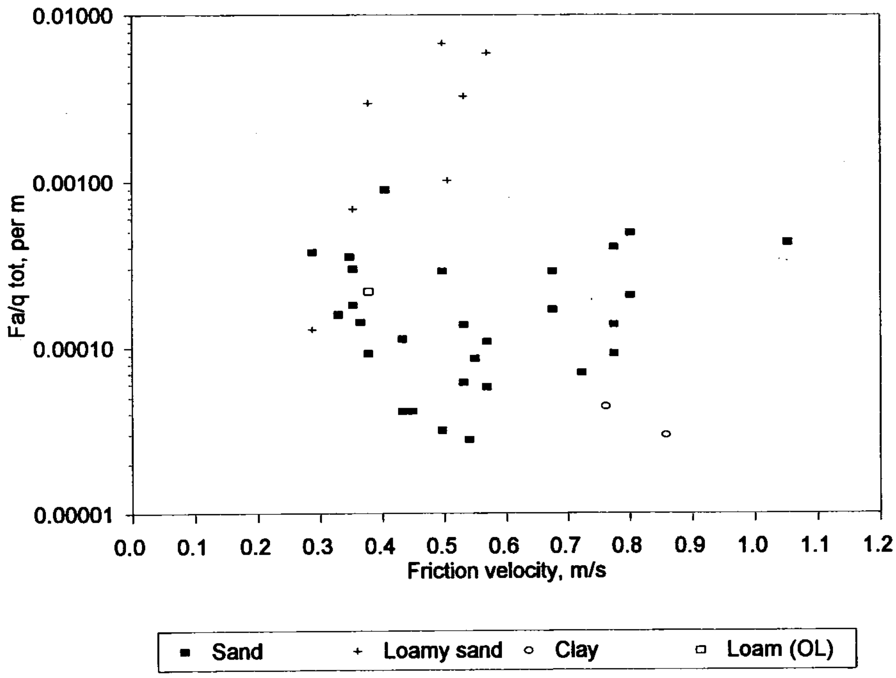

Figure 6: Ratio of vertical flux of PM_{10} to horizontal flux of total particle mass versus friction velocity for experiments for which different soil textures were found (after Gillette et al., 1997).

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 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 and 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. (1997) presented data shown in Figure 6 that shows F_a/q_{tot} results for the measurements made in Texas, and California, USA.

The F_a/q_{tot} data points for sandy soils appear in a group that does not seem to vary with friction velocity. The other textures of soils do not have enough points to affirm or deny a relationship with $u_\star$. Clay soils seem to produce less fine particle flux for a given total mass flux whereas loamy sand soils produce more; the loam-soil point is in the middle of the sand soil range. We hope to add more clay soils to our data base to confirm these apparent results. Examination of size distributions for q_{tot} values used in Figure 6 (Gillette and Chen, 1999) shows that the modes of the size distribution of saltating particles are within ± 20 µm for all the soils except the clay. The mode of the clay size distribution was roughly three times that for the other soils that corresponds to a mass of about nine times larger. The interpretation of Gillette and Chen of the data is that values are highest for clay, intermediate for sand and loam and lowest for loamy sand.

7 Proposed Prioritization of Mechanisms Responsible for Particle Emission Variability Observed at "Hot Spots"

Although much further research is required, I propose the following priority scheme for mechanisms (or absence of the mechanism) causing particle flux "hot spots". This list is made knowing that other mechanisms might be found to affect dust emissions and without building an integrated model for sensitivity testing. All these conditions should hold for the existence of a hot spot.

- High winds (above the threshold velocity).
- Lack of momentum partitioning, especially by vegetation, cobbles, boulders, or gravel. If there are particles $80 < d < 120$ µm then there will be a low threshold friction velocity for wind erosion.
- Lack of aggregation or crusting. Alternately, disturbance destroying aggregation.

- High particle availability.
- Owen effect (increase of drag coefficient with wind speed)
- Lack of trapping.
- Lack of cohesion by soil moisture.
- Low binding energies of suspended particles in the soil matrix.

8 Most Probable Locations and Conditions for a Wind-Erosion "Hot Spot"

The most probable locations and the conditions at these locations where the above mechanisms leading to wind erosion would be found are listed below. It should be remembered that these conditions derive from work done in "hot spots" in West Texas and Owens (Dry) Lake in California.

- Unvegetated and free of rocks, pebbles, cobbles. Vegetation, especially grass, leads to strong protection of the surface. Widely separated vegetation has more potential for dust emission than more continuous vegetation.
- Sandy sediments, typically laid down by fluvial deposit (ephemeral streams or flooding). Particle size, lack of crusting (aggregation) and smoothness lead to the lowest threshold friction velocities. If sand soil is not available, silty and clayey soil can be erodible by heavy disturbance.
- Long fetch (sediment lines up parallel to strong winds)–leading to the Owen Effect (nonlinear increase of flux with distance).
- Smooth ($z_0 < 0.1$ cm) – leading to low threshold friction velocity and lack of trapping. Slight rippling is advantageous to increase the drag coefficient.
- Disturbed sediment.
- Thick deposit–so that particle supply limitation does not take place.
- Located in a topography that converges winds.
- Dry. Lacking soil moisture and what is more important, soil crusting and aggregation.
- Strong meso-scale winds.

9 Conclusions

Consistently active dust producing areas "hot spots" are found within larger less active areas. The explanation of these "hot spots" on a geophysical level is lack of nonerodible roughness elements, low threshold friction velocity, and lack of aggregation or crusting. These conditions often apply in depositional environments (e.g., lake and river beds; deposits in regions of runoff from elevated terrains) and soils disturbed by man that are located in transitional arid regions. These environments are consistent with the "hot spots" discussed by J. Prospero.

Acknowledgement

The author would like to thank Prof. Joe Prospero for the inspiration to write this paper and his help during the writing. Mention of trade names or commercial products does not constitute endorsement or recommendation for use.

References

Abramowitz M. and Stegun I., 1964: Handbook of mathematical functions, National Bureau of Standards Applied Mathematics Series 55, U.S. Government Printing Office, Washington, D.C., 20402.

Arya S.P.S., 1975; A drag partition theory for determining the large-scale roughness parameter and wind stress on Arctic pack ice. J. Geophys. Res. **80**, 3447–3454.

Bisal F. and Hsieh J., 1966: Influence of moisture on erodibility of soil by wind. Soil Sci. **102**, 143–146.

Breuninger R.H., Gillette D.A., and Kihl R., 1989: Formation of wind-erodible aggregates for salty soils and soils with less than 50% sand composition in natural terrestrial environments. In: Leinen M. and Sarnthein M. (eds.), Paleoclimatology and Paleometeorology: Modern and Past Patterns of Global Atmospheric Transport. Kluwer Academic Publishers, 31–63.

Chatenet B., Marticorena B., Gomes L., and Bergametti G., 1996: Assessing the microped size distribution of desert soils erodible by wind. Sedimentology **43**, 901–911.

Chepil W.S., 1951: Properties of soil which influence wind erosion: IV. State of dry aggregate structure. Soil Sci. **72**, 387–401.

Chepil W.S., 1956: Influence of moisture on erodibility of soil by wind. Soil Sci. Soc. Am. Proc. **20**, 288–292.

Chepil W.S., 1958: Soil conditions that influence wind erosion. Technical Bulleting No. 1185, U.S.D.A. Washinton D.C., 40pp.

Chepil W.S. and Woodruff N.P., 1963: The physics of wind erosion and its control. In: Norman A.G. (ed.), Advances in Agronomy, Vol. 15, 1–301, Academic, San Diego, CA.

Cowherd C., Muleski G.E., Engelhart P., and Gillette D.A., 1984: Rapid assessment of exposure to particulate emissions from surface contamination sites. Midwest Research Institute Report of Project No. 7972-L. 425 Volker Boulevard, Kansas City, Missouri 64110.

Gillette D., 1981: Production of dust that may be carried great distances. In: Pe'we' T. (ed.), Desert dust: Origin, characteristics, and effect on man. Geological Society of America, Boulder, Colorado, 11–26.

Gillette D.A., 1988: Threshold friction velocities for dust production for agricultural soils. J. Geophys. Res. **93**, 12645–12662.

Gillette D., Adams J., Endo A., Smith D., and Kihl R., 1980: Threshold velocities for input of soil particles into the air by desert soils. J. Geophys. Res. **85**, 5621–5630.

Gillette D., Adams J., Muhs D., and Kihl R., 1982: Threshold friction velocities and rupture moduli for crusted desert soil for the input of soil particles into the air. J. Geophys. Res. **87**, 9003–9015.

Gillette D.A. and Chen W., 1999: The size effect in sandblasting of PM_{10} particles. In preparation.

Gillette D.A., Fryrear D.W., Gill T., Ley T., Cahill T.A., and Gearhart E.A., 1997: Relation of vertical flux of particles smaller than 10 μm to total aeolian horizontal mass flux at Owens Lake. J. Geophys. Res. **102**, 26009–26016.

Gillette D.A., Hardebeck E., and Parker J., 1997: Large-scale variability of wind erosion mass flux rates at Owens Lake, 2, Role of roughness change, particle limitation, change of threshold friction velocity, and the Owen effect. J. Geophys. Res. **102**, 25989–25998.

Gillette D.A. and Stockton P., 1989: The effect of nonerodible particles on wind erosion of erodible surfaces. J. Geophys. Res. **94**, 12885–12893.

Herman J.R., Bhartia P.K., Torres O., Hsu C., Seftor C., and Celarier E., 1997: Global distribution of UV-absorbing aerosols from Nimbus-7/TOMS data. J. Geophys. Res. **102** 16911–16922.

Husar R., Prospero J.M., and Stowe L.L., 1997: Characterization of tropospheric aerosols over the oceans with the NOAA advanced very high resolution radiometer optical thickness operational product. J. Geophys. Res. **102** 16889–16909.

Iversen J.D. and White B.R, 1982: Saltation threshold on Earth, Mars and Venus. Sedimentology **29**, 111–119.

Kaimal J. C. and Finnigan J.J., 1994: Atmospheric boundary layer flows, their structure and measurement. Oxford University Press, New York, 14–19.

Lancaster N. and Baas A., 1998: Influence of vegetation cover on sand transport by wind: field studies at Owens Lake, CA, Earth Surface Proc. and Land Farms, **23**, 69–82.

Marshall J.K.,, 1971: Drag measurements in roughness arrays of varying density and distribution. Agric. Meteorol. **8**, 269–292.

Marticorena B. and Bergametti G., 1995: Modeling the atmospheric dust cycle: 1. Design of a soil-derived dust emission scheme. J. Geophys. Res. **100**, 16415–16430.

Marticorena B., Bergametti G., Gillette D., and Belnap J., 1997: Factors controlling threshold friction velocity in semiarid and arid areas of the United States. J. Geophys. Res. **102**, 23277–23287.

Mc Kenna-Neuman C. and Nickling W.G., 1989: A theoretical and wind tunnel investigation of the effect of capillary water on the entrainment of sediment by wind. Can. J. Soil Sci. **69**, 79–96.

Musick H.B. and Gillette D.A., 1990: Field evaluation of relationships between a vegetationstructural parameter and sheltering against wind erosion. Land Degradation and Rehabilitation **2**, 87–94.

Owen P.R., 1964: Saltation of uniform sand grains in air. J. Fluid Mech. **20**, 225–242.

Priestley C.H.B., 1957: Turbulent transfer in the lower atmosphere. University of Chicago Press, 130pp.

Pye K., 1987: Aeolian dust and dust deposits. Academic Press, London, 334pp.

Raupach M.R., 1991: Saltation layers, vegetation canopies and roughness lengths. Acta Mechanica, I (suppl), 83–96.

Saleh A., and Fryrear D.W., 1995: Threshold wind velocities of wet soils as affected by wind blown sand. Soil Sci. **160**, 304–309.

Shao Y., Raupach M.R., and Findlater P.A., 1993: The effect of saltation bombardment on the entrainment of dust by wind. J. Geophys. Res. **98** 12719–12726.

Tapper N.J., 1991: Evidence for a mesoscale thermal circulation over dry salt lakes. Palaeogeography, Palaeoclimatology, Paleoecology **84**, 259–269.

Wolfe S.A. and Nickling, W.G., 1993: The protective role of sparse vegetation in wind erosion. Progress in Physical Geography, **17**, 50–68.