Saltating Sand Erodes Metastable Loess: Events in the Impact Zone

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Introduction: Scale

A saltating sand grain transfers energy from the wind to the ground. If the ground consists of the right materials in the correct condition then soil erosion can occur. The energy delivered to the point of impact causes the erosion, and this energy is required to do various things. The way in which the impact energy is distributed is considered in this paper, as is also the nature of the ground materials which suffers from erosion. What are the parameters that make it most vulnerable to wind erosion? Our aim is to highlight the metastability as a key ground property and to define the variables that control impact erosion. Also we will identify a parallel between soil structure collapse due to impact erosion events, and collapse due to hydrocompaction.

In their physical statistical approach to erosion, Hergarten and Neugebauer (1996) suggested that the evaluation of erosion models started in the 1950s with two complementary approaches:

1. Wischmeier et al (1958) developed the 'universal soil loss equation'; a completely empirical equation to calculate the soil erosion rate on a homogeneous area under given conditions.

2. Culling (1960) introduced a simple diffusion equation

    $$\partial/\partial t[H(\chi_1, \chi_2, t)] = D\Delta H(\chi_1, \chi_2, t)$$

    for the change of the surface height $H(\chi_1, \chi_2, t)$. This equation can be understood using very simple physical constraints. Although it may describe the smoothing of the land surface at large scales, this approach is neither able to predict erosion rates on smaller scales nor to explain the complex shapes and patterns produced by erosion. See Hergarten and Neugebauer (1996) for parameter identification and discussion.

Neither of these approaches brings us close to an actual examination of the soil erosion process. The scales are altogether too vast; the Culling approach operates at about 1 km^2, the Wischmeier et al system is perhaps at 1 m^2, whereas what is needed are studies at about 1 mm^2, at the interparticle, impact zone level. Studies at this level do not occur frequently; Smalley (1970) is possibly the only one date.

Material Available for Erosion

The soils of Western Kansas are currently classified as Udolls, damp Mollisols. They formed in loess material which was widely distributed over the Plains region during the Quaternary period. This silty material, which blew into position, could blow away again if it received enough kinetic energy from the wind; energy to break the interparticle bonds and lift the silt particles into the airstream. If the moisture regime locally falls from Udic to Xeric the tensile erosion can occur. The eroding system is essentially a dry, loosely packed silty soil. This is actually quite close to the model soil proposed by Hergarten and Neugebauer (1996) in which the pedosphere is assumed to consist of particles of fixed shape and size (they proposed cubes of size d). They separated the pedosphere

into two different layers: a lower layer formed by amalgamated particles, and an upper layer consisting of loose particles which can be moved if energy is added. The loose particles are assumed to be distributed randomly.

This Hergarten and Neugebauer model appears to be very suitable for a consideration of wind erosion in a loess soil system. In the short time since they proposed it several structural models for loess have been developed (Dibben et at 1996, Assallay et al 1997) each based on a Monte Carlo approach to soil structure and particle packing. This is soil structure studies at the single particle level, not the gross structure usually discussed in a pedological setting.

It is important to appreciate that there is a relatively limited range of material available for wind erosion. Silt erodes because there are some very efficient natural processes which produce particles at a mode size of around 20-30 μm (Jefferson et al 1997). If these particles were not available there would be next to no wind erosion. The ideal size for a wind eroded particle appears to be at about 80-100 μm; this is near enough to the supplied mode for large scale erosion to be feasible (and to occur). The particle types in soils are better defined than the grade scales and definitions allow for. There are natural geo-controls on the active clays, the inactive clays, the loess silts and the sands. Sand is not just an interval on a grade scale, it is the product of specific geological processes, and there is a strict lithogical control on size and shape (Smalley 1966, Krinsley & Smalley 1972).

The Impact Zone

Figure 1 shows the idealized view of the impact zone. A metastable loess soil is eroded by the impact of a sand grain. The energy lost by the sand grain during the impact event is available for erosion, with the addition of the energy contributed by the collapse of the metastable soil system.
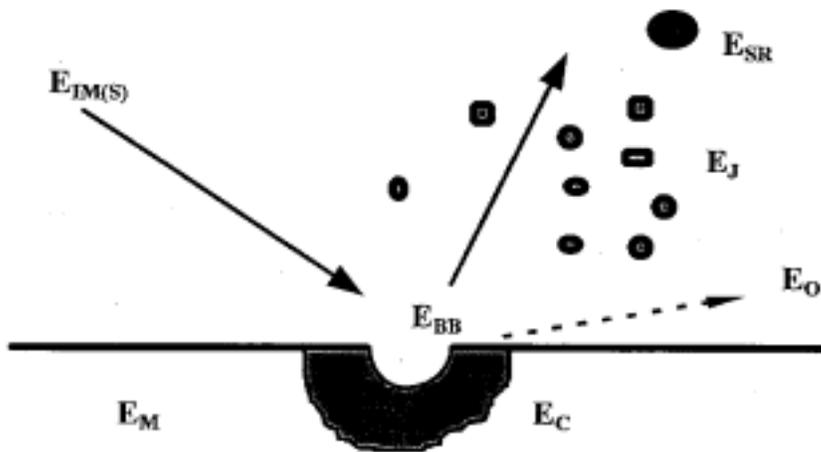

Figure 1.   Impact zone. A saltating sand grain strikes an ideal loess soil. The variables are described in the text, they indicate the distribution of energy in the system.

$$Eims(s) \; + \; Em \; - \; Esr \; = \; \text{erosion energy}$$

The erosion energy performs several tasks; it is required to break the interparticle bond to effect a bulk tensile failure in the critical erosion zone, to project, inject the eroded silt particles into the airstream so that they may be carried away in suspension. Enough energy will be retained by the impacting sand grain for it to carry on its way, and it also compact and disturbs the soil around the erosion zone. We identify two major variables on the energy dispersing side of the erosion equation, and propose a comparative equation to describe impact erosion by rain drops.

Two Simple Soil Erosion Equations

It should be possible to develop an equation which describes the key processes taking place in the critical impact zone where momentum is transferred from the impacting sand grain to the eroded silt particles. If the variables can be isolated and defined we can start to make quantitative estimates of their magnitude. In fact we should attempt to propose two equations; one for sand grain impact, and one for rain drop impact; comparisons may be mutually beneficial. We propose that energy transfer in the impact zone can be described by equation 1(see also Figure 1).

This represents the situation in which a saltatings sand grain strikes a loess soil surface:

$$Eim(s) \; + \; Em \; = \; Ebb \; + \; Ej \; + \; Esr \; + \; Ec \; + \; Eo \qquad\qquad (1)$$

Eim(s) is the impact energy delivered by the moving sand grain; this combines with Em, the metastable energy (potential energy) of the open structured loess soil system to cause disruption in the impact zone. Eim(s) delivers the trigger energy which makes the metastable stored energy available. The type of soil which suffers from wind erosion is likely to be a soil in which the Em value is significant.

The impact energy (plus metastable component) is dissipated into breaking interparticle bonds(Ebb); injecting eroded particles into the airstream(Ej); carrying the sand particle on its way(Esr) and disturbing and compacting the soil around the impact zone(Ec). It can be seen that Ebb and Ej are the important factors and the trade-off between these two factors affects the nature of the erosion. Esr and Ec are minor factors, in the consideration of the impact event, and there may be other minor factors which we have overlooked. A compromise between Ebb and Ej is needed for wind erosion to proceed. If the bonding is too strong a large Ebb requirement makes erosion impossible; if the particles are too large the Ej requirement also inhibits particle movement. There are doubtless other energy losses involved, but these are likely to be minor; some (a very small amount) of the impact energy will be converted into heat; these minor losses are included as Eo.

This compromise is what is studied in the series of ' flow/stick' experiments reported in this paper.

A similar equation might be written for soil erosion by rain drop impact. One should be careful about making too close a comparison between these two processes but there do appear to be some significant similarities. The nature of the Ej parameter may need special attention:

$$Eim(r) + Em = Ebb + Ej + Es + Esp + Ems + Ec + Eo \qquad (2)$$

The main inherent control factors are still Ebb and Ej but the minor factors are Es surface energy of newly created water droplets, Esp splash energy, Ems matric suction overcome, and Ec local compaction. Here the major factor trade-offs work somewhat differently. The particle is not lifted high into the eroding fluid stream so larger particles can be shifted; Ebb might be significant but the Ej requirement moves the erosive action to the larger (sand) particles. The bonding is short ranges but in the wet system this may not be significant. If there is very little Ebb requirement in the wet system Ej can move much larger particles. One would expect the mode size of rain drop eroded material to be much larger than that of sand grain eroded material. In the essentially dry wind eroded system Ebb will be substantial, and so will Ej.

Ej and Ebb

Both of these key parameters vary with particle size, in fact, in the short range bonded systems in which wind erosion occurs, particle size is the controlling factor.

$$Ej = \kappa_1, d^3 \qquad (3)$$

A simple cubic relationship links Ej and particle diameter. The injection energy required increases enormously as particle diameter increases: κ_1 is a scaling factor.

$$Ebb = \kappa_2 / d^3 \qquad (4)$$

An inverse cubic this time because the cohesion gets greater as the particle size decreases: κ_2 is a scaling factor. This is a simple short range cohesion that is being considered, no long range clay mineral type interactions are allowed.

With the rapid change of both parameters the minimum is very marked - as indeed Bagnold (1941 p.31) showed in his original and famous J-curve. Equation 4 relates to the tensile strength equation derived by Smalley (1970). This equation was designed to include all the relevant variables but in particular to include a factor for a failure volume, rather than a simple failure surface.

$$\sigma = (0.55B\rho kt) / d^3 \qquad (5)$$

Where σ is the tensile strength of the ideal loess soil system, B is the bond strength, ρ is the packing density, k is the coordination number and t is the failure volume. The coordination number k measures the number of soil particles touching any typical reference particle.

Equation 5, which is essentially the same as equation 4 shows the dependence of cohesion on particle size. Particle size is critical for these fine gained primary mineral particle systems.

The material concepts can be explored via the R-size diagram (see later) and tentative measurements can be made via the determination of the flow/stick transition.

The Flow/Stick Transition

Two major forces operate in a more or less dry granular system: the weight force and the cohesive force. If there is appreciable water in the system the whole picture becomes more complicated, and soil-water tension has to be considered. At very low water contents the inevitable moisture can be considered to be concentrated at the interparticle contacts and simple to enhance the cohesive forces.

In an ideal dry soil exposed to wind erosion the two forces tend to resist erosion, but in a definitely no-complementary manner. If the particle size in the system is increased weight forces oppose erosion, but cohesive forces in the critical impact zone are decreased. Small particles have increased cohesion but weigh less. The whole system becomes more cohesive as particle sizes is decreased, it transforms from a fee-flowing into a completely cohesive system. The flow/stick transition occurs, and the size at which the flow rate is a maximum should be the size of maximum danger for wind erosion; because at that particle size the erosion resisting forces are at a minimum.

The flow/stick transition was first measured by Smalley (1964) using a variant of the Bingham-Wikoff (1931) orifice flow apparatus (see Figure 2). The experiments produced a flow maximum (for crushed quartz sand particles) at about 150 μm particle diameter, with a complete flow stoppage at about 50 μm (see Figure 3). A more elaborate investigation was carried out by Jones and Pilpel (1966) and they showed that the Smalley results were influenced by the orifice size.

They used a range of orifice sizes (see Figure 3) and found a maximum flow rate at about 350 μm. Under ideal conditions this will be the particle size likely to be eroded in the impact erosion situation. The Jones-Pilpel experiments have been repeated at Loughborough and Nottingham Trent Universities and similar results obtained. If the cohesive forces were greater the optimum size for erosion would be greater. Farmer (1973) considering impact erosion by raindrops wrote "No fully satisfactory explanation has been advanced for the observed fact that detachability or rate of particle detachment is greatest in coarse and medium sand size material and reduced at either larger or smaller particle sizes". The explanation lies in the interplay of cohesive and weight forces, and the question is actually much easier to answer if transferred to a wind erosion scenario. In the dry systems where sand grains impact on arid ground the mode sized of eroded material will be smaller than Farmer's coarse or medium sand. In fact it will be appreciably smaller than the ideal 350 _m measured in the flow experiments because the particles have to be injected into the airstream and only weight is a factor here - so the mode will be pushed down towards the fine sand/course silt size range (material at around R = 1 on the R-size diagram, see Figure 4).
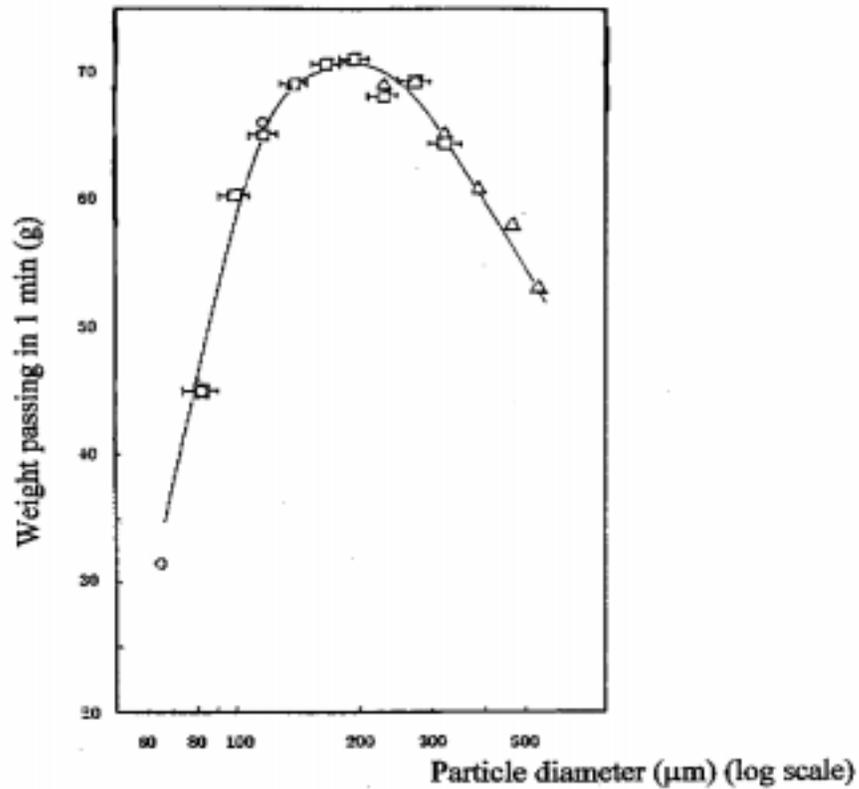
Figure 2.   Flow/stick transition. Flow rate vs. particle size for crushed quartz granular
material flowing through a 3 mm orifice; flow rate maximum at about 150 μm,
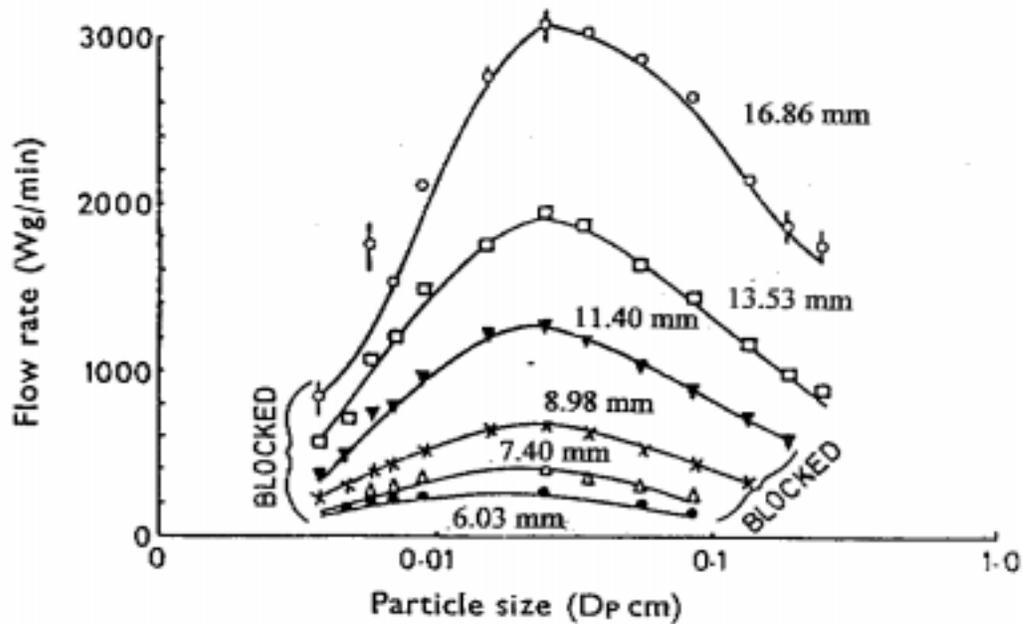flow stoppage at about 50 μm (after Smalley 1964).

Figure 3.   Flow/stick transition. Flow rate vs. particle size for granular magnesia flowing
through a variety of orifices; flow rate maximum at about 350 μm(after Jones
& Pilpel 1966).

The R-size Diagram

The R-size diagram supports the attempt to, in a general way, define the major types of particles found in soils. It plots (Figure 4) the bond/weight ratio R against the particle size and suggests a division of miner soil particles into five major divisions: A active clays, B inactive clays, C small inactive primary mineral particles, D silt and E sand.
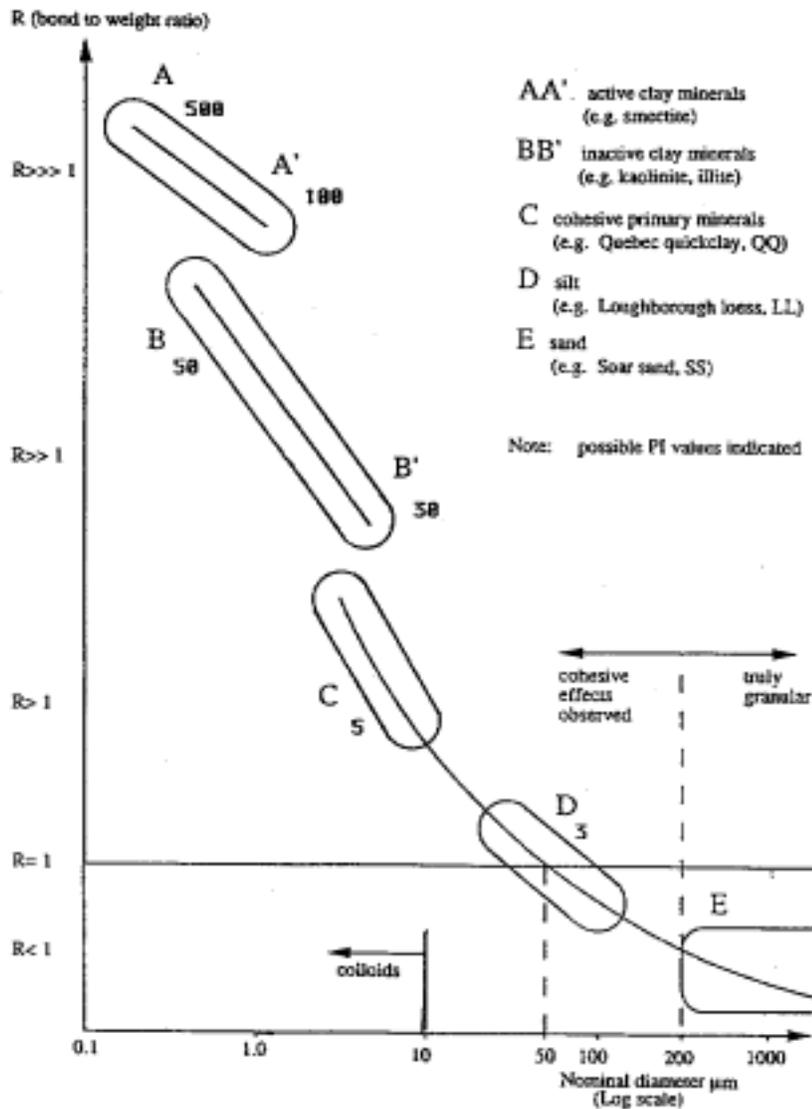

Figure 4.   R/size diagram. The bond/weight ratio R plotted against particle diameter for ideal granular soils. At R = 1 (at about 50 μm) the cohesive and the weight forces are equal. In this region we find some complex behaviour, and the loess soils subject to wind erosion.

The categories are related to production processes so that sand size is controlled by the eutectic reaction in granite (say 200-800 µm; 600 µm modal size, see Blatt 1987, Smalley 1966). At R = 1 the cohesive forces and the weight forces are just about in balance, in the soil engineering world this is the transition zone between cohesive and cohesionless soils (see Jefferson & Smalley 1995). The flow/stick studies put R = 1 at about 50 µm.

On the R-size diagram there is a fundamental transition between the charged particles (A and B) and the uncharged particles (C-E). The charged particles, the clay minerals, form soil systems with long range bonding, they have plasticity. The uncharged particles form short range bonded systems, they lack plasticity (see Cabrera & Smalley 1972); they may be cohesive, R somewhat greater than unity, but they lack plasticity. Increasing plasticity of the system is a sure way to reduce erosion.

The R-size diagram was designed to give a perspective on collapsing soils, but it has some relevance to wind erosion. The classic wind eroded material is type D silt.

Metastable Soils: Collapse and Erosion

The nature of the ground contributes to its behaviour under stress. It does appear that the soils which suffer most from wind erosion are metastable soils, soils which have the capacity for major deformation once a certain trigger energy level is exceeded; and the soils which suffer from wind erosion are essentially the same soils that suffer from structural collapse when loaded and wetted. A loess soil in Iowa may suffer from hydroconsolidation (if the clay mineral content is low enough: Handy 1973) and be located far enough to the east to perhaps avoid the problem of wind erosion.

A similar soil in W. Kansas, in periods of extreme aridity, could suffer from wind erosion (with the Em factor making a considerable contribution to the soil loss mechanism). The metastable soils (mostly loess based soils) have certain characteristics. They tend to be primary mineral soils, i.e. the clay mineral content tends to be low, and they have open structures with relatively loose packings of the soil particles. Short range bonds tend to predominate (Cabrera & Smalley 1973). In the loess soils the particle mineralogy is largely quartz and the shape of these loess particles is surprisingly flat. Rogers and Smalley (1993) have shown, using simple Monte Carlo methods, that the mode shape for a quartz silt loess soil particle should be fairly flat with an axial ratio of around 8:5:2. These flat particles can form open packing structures. The nature of these open packing structures is under investigation (see Assallay et al 1997, Dijkstra et al 1995, Rogers et al 1994) but the packing parameter has always been elusive and packing studies have made little progress in soil science or fine particle sedimentology.

It is just possible (see Jefferson et al 1997) that there is a lithological control on the size and nature of the quartz particles that comprise loess and related soils. The accumulation of crystal defects in granitic quartz may control the size and nature of quartz silt particles so that the open packing and interparticle bonding in an eroding soil are in fact influenced by the nature of the eutectic quartz crystals in the source granitic rocks. The predominance of these primary mineral

particles in metastable soils means that the interparticle bonding is of a short range nature, and that they are erodible.

Discussion and Proposals

The scheme of wind erosion discussed in this paper has many similarities to the 'sandblasting' approach of Alfaro (1997). He bombarded a kaolin clay target with mono-disperse quartz grains in a wind tunnel and studied the dust (silt)output. In R-size terms kaolin is a type B material and may not be the most realistic model for an eroding soil system. The high clay soils tend not to erode, but the kaolin does have a low clayeyness; the distinction between the Alfaro B-material and our D-material may not be that great. For the Alfaro aim of quantified wind erosion predictions to be reached a good understanding of events in the impact zone is probably required. We suggest that certain factors need to be considered, and certain questions asked about soils vulnerable to wind erosion:

1.   What is the nature of the primary mineral particles which are eroded? What is their shape and mineralogy?

2.   What packing structures are formed in erodible soils, what open initial structures? What is the level of metastability of the soil in present-day fields? How is impact erosion related to hydroconsolidation collapse?

3.   What is the nature of the interparticle bonding? How effective is the 'small clay' content of eroding/collapsing systems?

Some questions about the variables which operate in the impact zone:

1.   What are the key variables in the impact zone? How is the momentum carried forward from the impact point transfer?

2.   Are Ebb and Ej the key variables? Can others be safely neglected?

3.   How realistic is the model under consideration? Does it relate to the sandblasting approach of Alfaro, and how relevant is it to the case of impact erosion by raindrops? How does Ej function in the west system, and an we get from flow/stick observations a true measure of Ebb? More questions than answers.

Acknowledgments

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