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MEASUREMENT AND DATA ANALYSIS METHODS FOR FIELD-SCALE WIND EROSION STUDIES AND MODEL VALIDATION

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ABSTRACT

Accurate and reliable methods of measuring windblown sediment are needed to confirm, validate, and improve erosion models, assess the intensity of aeolian processes and related damage, determine the source of pollutants, and for other applications. This paper outlines important principles to consider in conducting field-scale wind erosion studies and proposes strategies of field data collection for use in model validation and development. Detailed discussions include consideration of field characteristics, sediment sampling, and meteorological stations. The field shape used in field-scale wind erosion research is generally a matter of preference and in many studies may not have practical significance. Maintaining a clear non-erodible boundary is necessary to accurately determine erosion fetch distance. A field length of about 300 m may be needed in many situations to approach transport capacity for saltation flux in bare agricultural fields. Field surface conditions affect the wind profile and other processes such as sediment emission, transport, and deposition and soil erodibility. Knowledge of the temporal variation in surface conditions is necessary to understand aeolian processes. Temporal soil properties that impact aeolian processes include surface roughness, dry aggregate size distribution, dry aggregate stability, and crust characteristics. Use of a portable 2 tall anemometer tower should be considered to quantify variability of friction velocity and aerodynamic roughness caused by surface conditions in field-scale studies. The types of samplers used for sampling aeolian sediment will vary depending upon the type of sediment to be measured. The Big Spring Number Eight (BSNE) and Modified Wilson and Cooke (MWAC) samplers appear to be the most popular for field studies of saltation. Suspension flux may be measured with commercially available instruments after modifications are made to ensure isokinetic conditions at high wind speeds. Meteorological measurements should include wind speed and direction, air temperature, solar radiation, relative humidity, rain amount, soil temperature and moisture. Careful consideration of the climatic, sediment, and soil surface characteristics observed in future field-scale wind erosion studies will ensure maximum use of the data collected.

KEY WORDS: wind erosion; saltation; aeolian processes; dust; sediment

INTRODUCTION

Accurate and reliable methods of measuring windblown sediment are needed to confirm, validate, and improve erosion models, assess the intensity of aeolian processes and related damage, determine the source of pollutants, and for other applications. The type of sampling apparatus and methods used in wind erosion (aeolian) field studies depend upon the specific objectives of the study. In this paper we will consider the field characteristics, aeolian sediment sampling devices, meteorological measurements, and selected data analysis methods for field-scale (paddock-scale) wind erosion studies used to validate and further develop wind erosion

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models. Studies at regional or larger scales may or may not require the same methods and will not be considered here.

Sampling windblown sediment in the field is not a new endeavor. Perhaps the earliest attempt was made in Western Australia in 1908 (Olsson-Seffer, 1908). In this study, the carrying capacity of the wind was estimated by measuring sediment collected at 2·0 cm intervals over a total height of 8·0 cm using five sheets of corrugated iron held together in a frame.

In later studies of dunes, Bagnold (1941) used a slotted collector with an opening 1·25 cm wide and 76 cm high to measure saltating grains and a buried ground trap to measure surface creep.

Chepil and Milne (1941) were probably the first to attempt to measure soil movement in agricultural fields. They used a slotted catcher to measure soil transported within 5 cm of the soil surface at various distances from the windward edge of the field. Few details about the wind or soil characteristics were observed. Chepil expanded this work on several major soil types from 1938 to 1944 using Bagnold-type sand catchers (Chepil, 1946) to relate surface saltation and creep flux to distance from the windward edge of the field (fetch). Many details of the experimental procedures are lacking. The study also included limited data on the size distribution of soil surface aggregates.

Chepil first attempted to estimate field soil loss attributed to wind erosion in 1960 (Chepil, 1960). Estimates of soil loss in mass per unit area were made by estimating the depth of soil loss from around wheat roots. The smallest depth resolution used to estimate soil loss was 1·25 cm. This paper presented tables relating estimated soil loss to field conditions during the measurement period of 1954 to 1956 in Kansas, USA. However, these estimates did not provide climatic and site characteristics often needed to test or develop detailed erosion models.

A detailed field procedure designed to validate erosion models, described by Fryrear *et al.* (1991), outlines field instrumentation and methods of data analysis. Recommendations included a circular 200 m diameter field with BSNE (Big Spring Number Eight) sampling masts (Figure 1) placed in a radial pattern within the circle and one mast outside the circle in a non-erodible location. The meteorological instruments included sensors for wind speed and direction, air temperature, solar radiation, soil temperature at one depth, rainfall, and relative humidity. A piezoelectric device called a SENSIT (Gillette and Stockton, 1986) was used to detect saltation particle impacts (Figure 2). A large number of sites have been instrumented in a similar fashion and used to develop or validate models such as the Wind Erosion Equation, Wind Erosion Prediction System, and Revised Wind Erosion Equation (Van Pelt *et al*., 2001, in press; Zobeck *et al*., 2001; Van Pelt and Zobeck, in press).

The Fryrear approach described above was specifically developed for agricultural land, particularly the most erosive fallow field conditions. Table I summarizes a wide variety of field-scale studies recently performed on agricultural land. In most studies, only one field was investigated at a location due to the large field size required or equipment limitations. A few studies have used paired plots to investigate spatial variation or treatment effects (Sterk and Spaan, 1997; Sterk *et al*., 1999; Gomes *et al*., 2003a).

A variety of erosion models are now available to estimate windblown sediment transport (Woodruff and Siddoway, 1965; Hagen, 1991; Marticorena and Bergametti, 1995; Shao *et al*., 1996; Potter *et al*., 1998; Fryrear *et al*., 1998; Alfaro and Gomes, 2001; Gregory and Darwish, 2001). Table II lists the climatic and soil variables needed for these models. Most models have input variables that include the need for data related to wind and other climatic variables, soil surface and near-surface properties, and vegetative properties. Validation of these and future models will require a systematic method of data collection based on common sampling principles and include methods of data collection for the input variables as well as sediment movement output variables. Although a variety of methods and instruments were employed in the studies described in Table I, not all studies included information on climatic, site, soil, and vegetative variables common to most models. The kind of information collected depended on the purpose of the study. Studies designed to validate erosion models (e.g. Zobeck *et al*., 2001) require much more data than an agronomic study in which conservation measures are tested (e.g. Bielders *et al*., 2000).

Even though a variety of methods and instruments have been used to investigate windblown sediment in agricultural fields, the selection of the instruments and the methods employed are governed by relatively few operating principles. This paper will outline important principles and factors to consider in conducting field-scale wind erosion studies and propose strategies of field data collection for use in model validation and development.

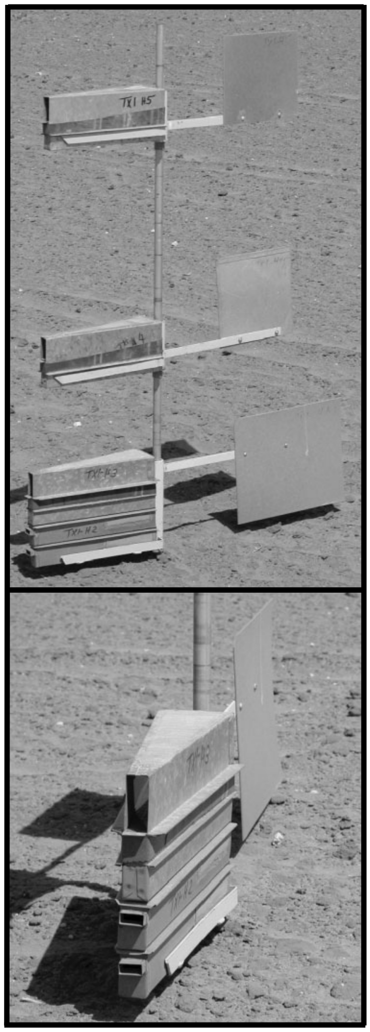

Figure 1. BSNE dust samplers. (Top) Samplers mounted at heights of 5, 10, 20, 50 and 100 cm. (Bottom) Close-up of bottom three samplers

This list is not intended to be exhaustive, but will include factors used by most models. The paper will discuss important field characteristics such as field shape, length, and surface conditions, sediment sampling, and meteorological station details. Consideration of slope, topography, and vegetation pose significant complexity to the challenge of measuring aeolian sediment and space limitations will not allow us to consider these here.

FIELD CHARACTERISTICS

Field geometry

Field shape. The field shape used in field-scale wind erosion research is important because it influences sampling design and data processing methods. Field shape is generally a matter of preference and in some studies may not have practical significance. Circular fields have the advantage of symmetry, which can simplify the calculation of fetch distances (Fryrear *et al*., 1991). Use of circular fields facilitates tests of the effects of ridges on wind erosion; tilling in a circular pattern allows testing of the effects of ridges perpendicular to the wind direction from all wind directions. The shape of the field is often imposed by local farming practices. For

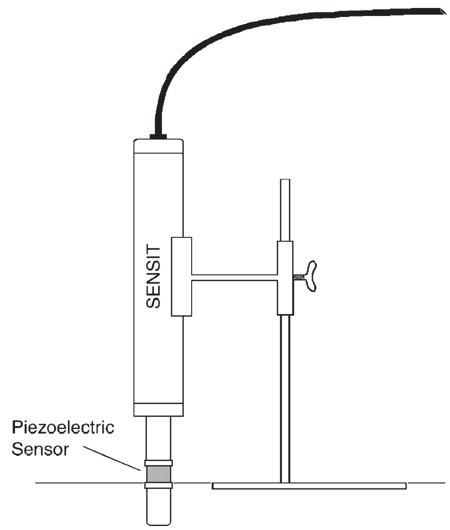

Figure 2. SENSIT particle impact sensor (adapted from Stout and Zobeck (1997) with permission of Blackwell Publishing)

example, rectangular or square fields are often the rule when working with farmers. Care should be taken to ensure that the long side of the field is oriented parallel to the dominant wind direction. This orientation is preferred to maximize the sampling fetch.

Field boundaries. Field boundaries are important in establishing a measurement or control area from which fluxes in and out of the study area are known. The precise nature of the boundary may limit experimental results and require specific sampling designs as discussed below.

Maintaining a clear, stabilized, non-erodible boundary is necessary to accurately determine erosion fetch distance. Tillage has been used to stabilize fields using tillage ridges in some studies (Fryrear *et al.*, 1991). However, the tillage ridges must be maintained in a rough, non-erodible condition so that previously eroded sediment is not blown back on the field when the wind direction changes. Stabilizing the edge of the field with flat residue or vegetation may be easier to maintain than tillage ridges, especially in the cases of sandy soils for which ridges collapse very quickly (Bielders *et al.*, 2000). If sediment is deposited upon the residue, application of additional residue will be necessary. Chemical soil stabilizers may be used if tillage and residue are not practical. Many commercial soil stabilizers are now available.

In many instances, it is impractical or not possible to accurately determine the non-erodible boundary. In these cases samplers should be installed to determine how much sediment crosses the field boundary into the study area. For example, in many areas such as the small-scale fields in Sahelian Africa, non-erodible boundaries are usually difficult to define. In such cases fetch may not be known, but it is important to know the input of sediment from upwind sources.

Samplers placed immediately upwind of the study field can be used to measure incoming sediment. If the incoming sediment is not accounted for in the analysis, the estimated soil losses may be incorrect. In addition, although net soil loss out of a study area may not be great, there may still be enough transport in such a field to damage crops and cause significant redistribution of soil material within the field, producing local degradation (at erosion spots) and enrichment (at sedimentation spots). The latter process is related to the high spatial variability in sediment mass fluxes to be discussed later.

Field length. Field length is important because it influences aeolian mass transport. Determination of the maximum sediment transport capacity of a field, needed to validate erosion in some models, requires a consideration of field length. Studies to evaluate wind erosion control practices must also consider field length when comparing treatments.

Saltation is a self-regulating phenomenon whereby saltation flux increases with fetch as the transport capacity of the wind, for a given field condition, is approached (Bagnold, 1941; Chepil and Milne, 1941; Owen, 1964;

Table I. Summary of recent field-scale wind erosion studies on agricultural land

Country	Citation	Field/plot size (m)	Saltation samplers*	Suspension samplers	Impact sensors	Soil properties	Cover type
Argentina	Buschiazzo *et al*., 1999	100 × 100	BSNE	None	Sensit	Texture, chem.†	Bare
China	Huang *et al*., 1997	Not given	Segmented samplers	None	None	Texture, organic matter chem.	Bare
Germany	Funk, 1995 Funk *et al*., in press	150 × 150	SUSTRA MWAC	None	Sensit	Texture	Bare
	Goossens and Gross, 2002	320 × 230	MWAC	MWAC	Saltiphone	Texture, organic matter	Cropped, plants <10 cm tall
Niger	Bielders *et al*., 2000	15 × 20	BSNE	None	Saltiphone	Texture, chem.	Cropped
	Bielders *et al*., 2002	Large	BSNE	None	None	Texture, chem.	Cropped
	Gomes *et al*., 2003a Rajot *et al*., 2002	100 × 100	BSNE	Low-Vol Active Suction	Saltiphone	Aggregate size istribution, crust type and thickness, loose erodible material	Bare
	Sterk and Spaan, 1997	55 × 70	MWAC	None	Saltiphone	Texture, aggregate size distribution	Bare and with residue
	Sterk and Stein, 1997	40 × 60	MWAC	None	None	Texture	Bare
Spain	Lopez *et al*., 1998	1·5 ha	None	Low Vol Active Suction	None	Texture, chem., wind erodible fraction, roughness, frontal area of roughness elements, loose erodible material, aggregate size distribution, bulk density	Bare
	Sterk *et al*., 1999 Gomes *et al*., 2003a Gomes *et al*., 2003b	135 × 180	MWAC	None Low Vol Active Suction	Saltiphone	Texture, chem., wind erodible fraction, roughness, frontal area of roughness elements, crust thickness and strength	Bare
US	Stetler *et al*., 1994 Stetler and Saxton, 1996	1600 × 1000	BSNE	High Vol. and Low Vol. Active Suction	Sensit	Texture, organic matter	Bare
	Stout and Zobeck, 1996	360 × 250	BSNE	None	Sensit	Texture, organic matter, roughness	Bare

* BSNE, Big Spring Number Eight sampler; SUSTRA, Suspended Sediment Trap; MWAC, Modified Wilson and Cooke sampler
† Chem, chemical properties of soil usually include electrical conductivity, carbonate, and pH.

Table II. Climatic and soils variables used by selected wind erosion models

Model*	Citation	Climate variables	Roughness	Aggregation†	Crust‡	Soil
WEQ	Woodruff and Siddoway, 1965; Soil Survey Staff, 1988	Precipitation, potential evapotranspiration, mean monthly wind direction, preponderance and % erosive energy, irrigation dates	Ridge height, ridge spacing, standing and flat biomass	Fraction <0·84 µm	Presence or absence	Sand, silt, clay, organic matter, and calcium carbonate content
RWEQ	Fryrear *et al.*, 1998	Wind speed, precipitation, air temperature, solar radiation, irrigation dates	Ridge height, ridge spacing, random roughness, ridges orientation, standing and flat biomass	Function of sand, silt, clay, organic matter and calcium carbonate content	Clay and organic matter content	Sand, silt, clay, organic matter and calcium carbonate content
WESS	Potter *et al.*, 1998	Wind speed, wind speed perturbation factor	Ridge interval, roughness factor, standing and flat biomass	Soil erodibility as a function of soil texture	Not considered	Bulk density, erodible particle diameter, soil water content and drying rate
WEPS	Hagen, 1991	Air density, wind direction, mean daily wind speed in specified intervals per day or Weibull equation factors and fraction calm	Ridge height, ridge spacing, random roughness, ridge width, ridge orientation, dyke spacing, standing and flat biomass	Aggregate density, aggregate stability, GMD, GSD, min. aggregate size, max. aggregate size	Crust fraction, crust thickness, fraction LEM on crust, mass of LEM on crust, crust density, crust stability	Soil layers number, following by soil layer: thickness, bulk density, sand, silt, clay fraction, rock volume, wilting point water content, water content
TEAM	Gregory and Darwish, 2001	Wind speed and height of wind speed measurement, relative humidity	Function of surface covered with aggregates and average diameter of aggregate	Not considered	Not considered	Sand, silt, clay content, mean diameter of non-clay component, d_{50}, and mean diameter of particle between d_{50} and 1·0 µm
DPM	Marticorena and Bergametti, 1995; Alfaro and Gomes, 2001	Wind speed and air temperature profiles	Aerodynamic roughness length	Not considered	Not considered	Dry soil texture, soil water content
WEAM	Shao *et al.*, 1996	Wind speed, precipitation, evaporation	Aerodynamic roughness length, plant and soil frontal area index	Not considered	Not considered	Coarse, medium and fine sand content

* WEQ, wind erosion equation; RWEQ, revised wind erosion equation; WESS, wind erosion stochastic simulator; WEPS, wind erosion prediction system; TEAM; Texas Tech erosion analysis model; DPM, dust production model; WEAM, wind erosion assessment model

† GMD, geometric mean diameter; GSD, geometric standard deviation.

‡ LEM, Loose erodible material.

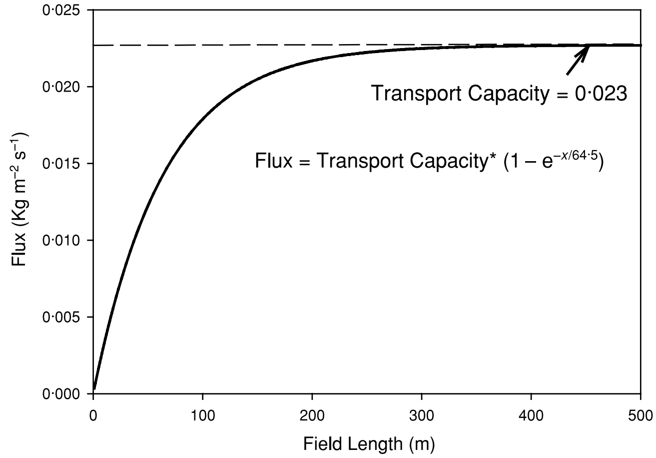

Figure 3. Saltation transport measured at a height of 0·15 m in an eroding fine sandy loam on 2 March 1988 as described by Stout (1990)

Stout, 1990; Gillette *et al*., 1996). The rate at which mass transport increases with distance is a function of field characteristics. A simple equation to describe the increase of saltation flux with distance at a specific height above the soil surface has been described by Stout (1990) as:

$$f(x,z) = f_{mx}(z)(1 - e^{-x/b(z)}) \tag{1}$$

where $f(x,z)$ is the saltation flux at distance x and height z, $f_{mx}(z)$ is the maximum saltation transport capacity at height z, and $b(z)$ is a length scale parameter that also varies with height. The b parameter will vary with surface characteristics such as soil texture, roughness, residue cover, etc. Note that where $x = b$, Equation 1 reduces to 0·632, indicating that b represents the distance at which mass flux attains a value of 63·2 per cent of the transport capacity, f_{mx}. About 98 per cent of the transport capacity is reached at a distance of 4b. Figure 3 illustrates the saltation flux measured at a height of 0·15 m in an eroding fine sandy loam (Stout, 1990). As transport capacity is attained, saltation flux remains relatively unchanged because the momentum extracted from the wind by particles already in flow limits the amount of mass that may be entrained and, thus, deposition and erosion of sediment are balanced.

The transport capacity and distance required to reach the transport capacity vary with surface roughness, amount of erodible material, vegetation or other cover, and other factors. A bare, smooth, flat field with unstructured loose but uniform soil would present the most erodible condition. Any changes in any of these factors could cause differences in saltation flux due to differences in entrainment, transport, or deposition of sediment (Stout and Zobeck, 1996). When any of these factors change, they should be carefully documented. For most model validation purposes, the field should be 'homogeneous'. It may be better to evaluate a small homogeneous field than a large heterogeneous field with great variation in surface properties. Unfortunately, most agricultural fields are characterized by spatial variability in the many factors that determine the wind erodibility of the soil, and thus significant spatial variation in saltation flux is often observed.

A field length of approximately 300 m is needed in many bare agricultural fields to approach saltation transport capacity. Analysis of data from a 200 m diameter bare, flat, fine sandy loam field in Big Spring, Texas, suggests maximum saltation flux was not reached for most erosion events. Saltation flux was near maximum at a distance of 250 m in a bare, fine sandy loam soil with small tillage ridges (Stout and Zobeck, 1996). Smaller fields may be adequate in sandier soils, although in a study of a sandy soil in Niger (greater than 90 per cent sand), the transport capacity was not reached at 80 m (Bielders *et al*., 2002).

In studies to evaluate wind erosion control practices, smaller fields are often used to compare the saltation transport among various erosion control alternatives. Usually transport capacity is not attained in these smaller fields. It is most important in these studies to ensure that the field lengths used to compare among treatments are the same or have been scaled to the same distance among treatments. This is necessary because, as discussed above, until the transport capacity is reached the saltation transport increases with field length. Scaling to the same distance in each treatment is simplified if the sampling distance is within the linear portion of the saltation transport curve as illustrated in Figure 3. Additional details of sampler location to evaluate erosion control practices are described later in this paper.

Suspended particles never reach a stabilized maximum in fields but continue to be produced, transported, and dispersed as long as saltation is occurring at the soil surface. For example, the horizontal mass flux of suspended particles continued to increase at distances greater than 350 m in a bare, fine sandy loam soil with small tillage ridges (Stout and Zobeck, 1996) and at distances greater than 1560 m for a salty dry playa lake surface (Gillette *et al*., 1997). Nevertheless, it is useful to determine the vertical suspension flux, which is a function of horizontal saltation flux (Gillette, 1977; Shao *et al*., 1996; Alfaro and Gomes, 2001). Although the relationship of vertical suspension flux and saltation flux depends on the fetch length and can theoretically be measured at many places in the field, we recommended making measurements where saltation flux has reached a maximum, if possible.

Field surface conditions

A variety of surface conditions or factors should be considered in field-scale wind erosion studies. Table II lists the soil surface properties used in selected wind erosion models now in use. Some properties such as soil texture, organic matter, and calcium carbonate content affect the erodibility of soils but are inherent properties that change very slowly under natural conditions. They should be documented for each study but usually need to be measured only once. Other properties, such as surface roughness, are temporal and change rapidly in response to climatic conditions or management practices. Temporal soil properties known to be important in wind erosion processes should be measured whenever significant change occurs during the sampling period. Details of the soil properties affecting WEPS have been provided by Zobeck (1991b). Other models may require different representations of the same properties or require different properties (Table II).

Surface soil wetness. The wetness or moisture content of the surface layer of particles, about 2 or 3 mm in depth, has a profound effect on the wind erodibility of unconsolidated sediment. Many studies have shown that the intergranular cohesion associated with moisture increases the threshold velocity needed for particle movement (Chepil, 1956; Bisal and Hsieh, 1966; Azizov, 1977; McKenna-Neuman and Nickling, 1989; Kroon and Hoekstra, 1990; Seleh and Fryrear, 1995; Chen *et al*., 1996). A review of the effects of surface moisture content on aeolian sand transport has been provided by Namikas and Sherman (1995).

Traditionally, water content has been expressed as the gravimetric (mass) soil water content (Θ_m), the ratio of the mass of water present in the sample to the mass of oven-dried (dried to constant weight at 105 °C) soil (Topp, 1993). However, the water content may also be expressed on a volumetric basis (Θ_v), as the volume of water present in a volume of soil. The water content expressed on a volume basis is calculated as the product of Θ_m and the ratio of the bulk density of the soil to the density of water. The distinction is important and should be clearly indicated when reporting results.

Although many studies of the effect of sediment water content on aeolian transport have related transport to the water content on a mass basis, the water content on a mass basis does not indicate how tightly the water is held by the soil, which is referred to as the water potential. The water potential may be thought of as the amount of work required to remove water from the system (Livingston, 1993). The curve describing the relationship of water content on a volume basis and the water potential is called the soil water release or soil moisture characteristic curve. In practice, water release curves are uniquely related to a soil and are most closely associated with soil texture and structure. As a result, soils with different particle size distributions will not have the same soil water potential at the same water content. For example, the water in a clay soil with 15 per cent water content will be bound much more tightly than the water in a fine sandy loam soil also at 15 per cent water. The importance of this relationship is underscored by McKenna-Neuman and Nickling (1989), who showed that the entrainment of sediment is theoretically and physically related to soil water potential. Field measurements of soil water should be related to the water potential to have wide application. Relating water content on a volume or mass basis to water potential is typically done using laboratory pressure plate chambers that maintain a constant pressure on a wet sample. The water content is measured on a volume or mass basis on samples that have been brought to the equilibrium moisture level at a known pressure (Klute, 1986b).

There are many methods to measure the water content of soils in the field (Klute, 1986a; Carter, 1993). Most methods require installation of a sensor within the soil matrix, or in the case of gravimetric measurements, require the destruction of the sample. However, since wind erosion is a surface phenomenon, only the first few millimetres of particles are immediately affected and must be observed. Measurements of soil wetness collected even a few centimetres below the soil surface are not directly related to aeolian entrainment and transport. The water potential of soil materials dry enough to be entrained by wind is best measured in small chambers by thermocouple psychrometry (Rawlins and Campbell, 1986). It is not currently possible to use thermocouple psychrometers to measure the soil surface 3 mm *in situ*.

Of the many methods to measure soil water content, only gravimetric water content measurements have been used to measure the 0–3 mm near-surface layer. Gravimetric sampling is the standard method of measuring water content of the near-surface layer in wind tunnel studies of the effect of soil wetness on entrainment. A very simple, easily constructed, and inexpensive field soil sampler to determine the soil-water content distribution in layers as thin as 1 mm has been described by Reginato (1975). However, gravimetric sampling is time-consuming and requires modification of the eroding surface.

There are no thoroughly tested, commercially available instruments to measure the *in situ* water content of this very thin (0–3 mm or so) surface layer. One recent commercially available instrument (Pier Electronic, GmbH, Wallau, Germany; http://www.pierelectronic.com/index.html) uses surface reflectance measured with an infrared photometer to estimate moisture content. This device requires calibration, but has been used with success in evaporation studies (Funk and Frielinghaus, 1997). Unfortunately, this instrument may have limited application because the depth of penetration is dependent on the moisture content. Additional research is needed to develop an inexpensive, real-time continuous reading of the field water potential of the surface 3 mm.

Although aeolian entrainment of sediment is clearly related to soil wetness, establishing the level of wetness in the field by any method is difficult due to high spatial and temporal variability. Indeed, the near-surface soil moisture varies considerably with depth over very short time periods and the soil is not uniformly moist across the landscape in the field. We have observed aeolian transport in a tilled fine sandy loam soil shortly after a thunderstorm passed the field, when standing water was still visible. In this case, a very thin layer of loose erodible material deposited upon the ridges quickly dried and began blowing while the furrows still held ponded water. A recent study of aeolian transport across a beach (Jackson and Nordstrom, 1997) describes the highly variable nature of the beach surface water content in space and time. Gravimetric samples of the 0–5 mm depth were made at several locations during a high wind event to document the temporal and spatial variability of surface wetness. Future studies of field wind erosion that seek to account for the effect of soil wetness should address the spatial and temporal nature of soil wetness affecting the eroding surface.

Surface roughness. Soil surface roughness (SSR) affects the wind stream by modifying parameters used to describe the wind stream such as aerodynamic roughness and friction velocity. These parameters will be described in more detail later in this paper. At the field scale, SSR or microrelief refers to the small-scale surface features produced by tillage ridges, clods, rocks, or other surface features, excluding vegetation or other large obstructions present on the soil. Surface roughness can be expressed in a number of ways and models vary in the types of information they use to relate the effects of roughness with wind erosion. While some models use aerodynamic roughness and friction velocity exclusively, others also require specific details of SSR (Table II).

Tillage produces oriented roughness called ridges (created by tillage tools in the direction of tillage) and random roughness produced by the random orientation of soil aggregates and clods. The roughness produced by tillage or other mechanical disturbances is also called microrelief. Wind erosion is affected by both oriented and random roughness (Armbrust *et al.*, 1964; Fryrear, 1984). Oriented roughness can be described by measuring tillage ridge height and spacing. The ridge-to-height ratio is used in the Wind Erosion Equation (WEQ) (Woodruff and Siddoway, 1965).

Random roughness (RR) is usually expressed as the standard deviation of height elevations after removing the effects of surface slope and oriented roughness (Currence and Lovely, 1970). Surface pin meters are often used to collect elevation measurements at 5 cm intervals over a 1 m^2 area. More recently, laser profile scanners have been used to automatically collect roughness data (Figure 4). The Revised Wind Erosion Equation (RWEQ) (Fryrear *et al.*, 1998) and the Wind Erosion Prediction System (WEPS) (Hagen, 1991) are examples of models that use random roughness.

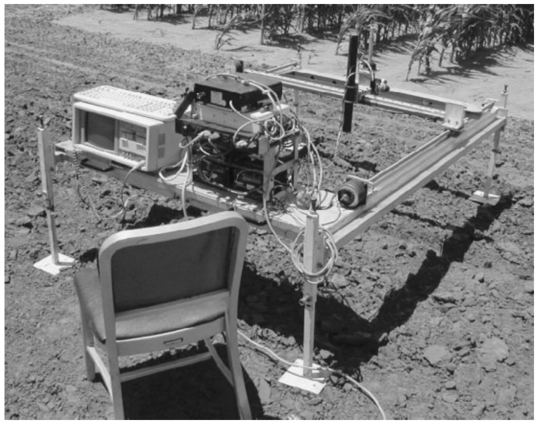

Figure 4. Automated laser-based soil surface roughness meter

A simple method using a 1 m long chain, called the chain method, has been developed to estimate random roughness (Saleh, 1993). The chain is laid on the ground and the shortened length has been related to pin-meter measurements of random roughness. This method may be difficult to use in tilled fragile soils since the chain must remain on the soil surface.

A new microrelief index has been developed for use in WEPS that can be used to estimate the fraction of soil surface susceptible to abrasion (Potter *et al.*, 1990). To resolve the index, surface elevation measurements collected on a 1 m^2 area are described by a two-parameter Weibull function:

$$SF = 1 - \mathrm{EXP}(-(SA/B)^C) \tag{2}$$

where *SF* if the surface fraction of observation points having a shelter angle less than or equal to a given shelter angle, *SA* is the given shelter angle, and the *B* scale parameter and *C* shape parameter are estimated by least-squares non-linear regression. The shelter angle is defined as the minimum angle from horizontal that a saltating particle must descend in order to strike a given location. Further details of the method of calculating this index are provided by Potter *et al.* (1990). This method requires elevation grid data and is sensitive to rainfall amount and intensity, tillage implement, and soil type (Zobeck and Popham, 1997, 2001).

The WEAM model of Shao *et al.* (1996) uses the non-erodible soil frontal area index of particles >0·85 mm diameter as a measure of SSR (Table II) defined as:

$$\mathrm{Index} = nbh/s \tag{3}$$

where n is the number of roughness elements on the ground area s, and b and h are the characteristic width and height of the elements, respectively. Lopez *et al.* (1998) used this index in a study of wind erosion in Spain (Table I).

Dry aggregate size distribution. Dry aggregate size distribution (DASD) refers to the relative amounts of air-dry aggregates or clods, on a mass basis, by size class, present on the soil surface. The type of data used to represent the dry aggregate size distribution depends on the model used. A model developed by Marticorena and Bergametti (1995) focusing on desert soil mobilization under natural, usually non-crusted, conditions uses threshold friction velocity computations that are dependent on the microped size distributions of erodible soil particles (Sørensen, 1985; Li and Martz, 1994).

Chatenet *et al.* (1996) have proposed a sampling and analysis method for determining the dry size distribution of loose microped material. Approximately 0·5 to 1 kg samples are oven-dried at 105 °C for 24 h, followed by cooling in a desiccator. The fraction greater than 2 mm is separated by slow hand-sieving. The remaining sample is divided into 70 g subsamples that are sieved for 8 min on an eight-sieve Fristch vibratory system set at a vibration intensity of 5·5 (Chatenet *et al*., 1996). The DASD is described as the geometric mean and geometric standard deviation of a combination of log normal distributions representing the mean mass collected on the sieves. Caution should be exercised when using the geometric mean and standard deviation because these parameters are greatly affected by specification of the smallest sieve size and sieving method (using amount passing versus amount retained on each sieve; Zobeck *et al*., 2003).

Other models such as WEPS and RWEQ focus on soil mobilization in disturbed agricultural fields and so consider the entire distribution of soil aggregates and clods. For these models, approximately 5 kg samples are collected from the upper 5 cm of the disturbed (tilled) soil surface, air-dried and sieved using a rotary sieve (Chepil, 1962). The rotary sieve is a sieving device that has been used in the USA to determine the erodible fraction (mass fraction of particles <0·84 mm diameter) of surface soils for many years. If tillage ridges are present, samples should represent the entire bed form from the tillage ridge to the furrow bottom. Recent analyses of over 5400 distributions have demonstrated that the Weibull distribution is the most accurate and precise distribution to describe DASDs (Zobeck *et al*., 2003). The Weibull distribution used to describe soil DASDs is of the form:

$$\text{Fraction at size } X = \frac{M(x < X)}{M_T} = 1 - e^{-(X/b)^c} \qquad (4)$$

where $M(x < X)$ is the sample mass x passing sieve opening diameter X, M_T is the total sample mass, the b parameter is a scale factor and the c parameter is a shape factor. The geometric mean diameter is estimated as the size at 0·50 fraction passing (d_{50}). The erodible fraction is determined by solving for Equation 4 after substituting 0·84 mm for X.

Dry aggregate stability. Dry aggregate stability refers to the resistance of soil aggregates to breakdown from physical forces and is a measure of the strength of the binding agents within aggregates (Skidmore and Powers, 1982). During wind erosion, soil aggregates are subject to bombardment by saltating grains that impart physical forces upon impact. The amount of erodible sediment dislodged from aggregates is related to how well they resist abrasion (Hagen, 1984). Although a variety of methods have been developed to estimate dry aggregate stability (Chepil, 1951; Skidmore and Layton, 1988), few are used in modelling the erosion of soil by wind.

A notable exception is the WEPS model which uses a dry aggregate stability index called the crushing energy. The crushing energy is a measure of the amount of energy per unit mass needed to crush an aggregate of a specified size to a specified end point (Skidmore and Powers, 1982). A device called a crushing energy meter has been developed specifically to measure crushing energy (Boyd *et al*., 1983; Hagen *et al*., 1995). The soils are sieved in the field to obtain aggregates approximately 15 mm diameter. Due to the large variation in crushing energy values for individual aggregates, at least 15 aggregates are collected from each location.

Crust Characteristics. In agricultural fields, crusts form in unconsolidated soils during rainfall and are affected by factors such as soil aggregation and roughness (Zobeck and Popham, 1992). Crusts are more compact and mechanically stable than the soil below (Chepil, 1958) and have a great impact on the erodibility of soil by wind. The effects of crusts on wind erosion have been measured in wind tunnels (Zobeck, 1991a; Hagen *et al*., 1992; Rice *et al*., 1996, 1997) and in the field (Sterk *et al*., 1999; Gomes *et al*., 2003b; Rajot *et al*., 2003). Wind tunnel studies of 14 soils found unconsolidated, loose soil to be from 40 to 70 times as erodible as the same crusted soils (Zobeck, 1991a). However, few models use the presence or absence of a crust in erosion estimates (Table II). Adjustments to the soil erodible fraction and estimated annual soil loss are made in WEQ when a crust is present (Soil Survey Staff, 1988, table 502-2). Crust thickness, stability, fraction of cover, and loose erodible material lying on the crust are used in WEPS (Zobeck, 1991b).

Since the effect of crust characteristics on aeolian processes is poorly understood, measurement methods have not been widely accepted. For example, crust thickness, although conceptually simple, is difficult to determine. Most agree that the crust includes the surface skin-seal of fine material and the deeper washed-in region immediately below the skin-seal (McIntyre, 1958). The thickness of a crust is measured in the field using a simple ruler. The difficulty comes in identifying the somewhat arbitrary separation of the washed-in region and unconsolidated material beneath. The situation is further complicated when larger clods intrude into the washed-in zone of the crust.

Figure 5. Pocket penetrometer showing: A, 6 mm diameter foot; B, strength level indicator; C, location of spring within penetrometer body

Perhaps the most convenient method to measure crust characteristics is to use a line-transect, making crust observations at regular intervals along the transect (Zobeck and Popham, 1992; Rajot *et al*., 2003). Measurement of crust thickness, stability and amount of loose erodible material lying on the crust may be made at each point.

Crust stability or strength, in the context of wind erosion, refers to the ability of crusted soils to withstand the abrasive action of saltating sand grains during wind erosion events (Zobeck and Popham, 1992). A commercially obtainable spring-loaded pocket penetrometer (Figure 5) is available to make many rapid measurements of crust stability in the field. The pocket penetrometer is used to measure the force necessary to rupture the crust. As the 6 mm diameter foot (A in Figure 5) is pushed into the soil, a strength level indicator (B Figure 5) is moved down the arbitrary scale. The level indicator records the maximum strength as the foot springs back to the starting position when the crust fails. The force can be determined by calibrating the arbitrary strength reading on the penetrometer to measurements made on a balance. In soils with very weak crusts, the spring within the penetrometer (C in Figure 5) may need to be replaced with a weaker spring. However, direct application of this technique for evaluating crust erodibility by saltating grains is questionable. Rice *et al*. (1997) described the use of a flat-tipped 0·6 mm diameter penetrometer and a flat-ended cylindrical punch with inner and outer diameters of 5 and 6 mm, respectively, to estimate crust strength. They suggested that the small penetrometer gave results that can be used to characterize surface erodibility to saltating particles. They determined that the punch results would be unsuitable to estimate erodibility, as would other strength tests that are on too large a scale.

The amount of loose erodible material present on the crusted surface is related to the wind erosion potential. These materials act to abrade the soil surface, causing further erosion. Erosion in some areas is limited by the amount of erodible material available to abrade a crusted surface. Scientists working on WEPS have measured the mass per unit area of this loose material using a specially designed field-portable vacuum system (Zobeck, 1989). A small portable commercial vacuum cleaner will also work in most situations where a power source is available. A simple and rapid method was recently used to measure the depth of this loose erodible material along a transect in Niger (Rajot *et al*., 2003). In this study, the loose erodible material depth was measured by observing the depth that a needle penetrated after stopping at the crust surface.

Non-erodible surface cover. Non-erodible surface cover includes any material lying on the soil surface, protecting it from the force of the wind and impact of saltating grains. In most models, non-erodible surface cover includes coarse rock fragments and flat-lying plant residue. Generally, there is an exponential decrease in soil loss with increasing surface cover (Bilbro and Fryrear, 1995). Non-erodible surface cover can be determined using line transects (Laflen *et al*., 1981) or by counting points touching cover elements on a grid overlaid on a nadir-view photograph of the field.

Sampling design. The sampling design used to describe field characteristics will vary with landscape complexity, purpose of the study, available time, and monetary constraints. A wide variety of methods are available to describe the spatial variability of field characteristics. Detailed discussion of sampling designs is beyond the scope of this paper. Papers focusing on soil spatial variability were presented at an international workshop in 1984 (Nielsen and Bouma, 1985). Simple line transects, as described above, can be used to determine the fraction of cover of clods, vegetation, crust, loose erodible material or other features needed for erosion models such as WEPS. More sophisticated geostatistical approaches may be used to provide maps of field characteristics. Maps of field surface characteristic may be used in field studies to evaluate the effect of a characteristic on a downwind sampling location.

Field characteristics basic operating principles and strategies

Although field shape is generally a matter of preference, some shapes have definite advantages. Circular fields can simplify the calculation of fetch when samplers are placed symmetrically in the field and tillage in a circular pattern allows testing of the effects of ridges from any direction.

Clear identification of a non-erodible boundary is necessary for the determination of distance to reach maximum saltation transport. A field length of about 300 m is necessary to approach maximum saltation transport capacity in most bare agricultural fields. Any change in surface conditions will usually alter the soil erodibility and thus, the maximum transport capacity and fetch needed to reach maximum transport. Suspended particles never reach a stabilized maximum in fields, but continue to be produced, transported, and dispersed as long as saltation is occurring at the soil surface.

Knowledge of spatial and temporal variation in surface field conditions is necessary to understand aeolian processes. Field surface characteristics that may affect aeolian processes include surface wetness, roughness, dry aggregates size distribution, dry aggregate stability, crust characteristics, and non-erodible surface cover. Baseline values for these characteristics should be established at the start of data collection and measured as often as necessary when significant change has occurred in surface conditions.

AEOLIAN SEDIMENT SAMPLING

Sampler design and efficiency

The types of samplers used for sampling aeolian sediment will vary depending upon the type of sediment to be measured. Aeolian sediment samplers may use a passive or active sampling process (Zobeck, 2002). Passive samplers rely on ambient wind conditions when collecting samples. Active samplers rely on some type of suction provided by a vacuum pump to draw a known volume of air and particles into the device. Active samplers are often used to collect suspended dust (defined as particles <60 μm diameter according to Bagnold (1941) or <20 μm diameter according to Gillette (1977)). Ideally, all samplers should be isokinetic, whereby the same wind speed is maintained through the orifice of the samplers as the ambient wind at the same height. A third type of sampler that may be more appropriately termed a 'sensor', detects the presence of saltating grains by recording impacts upon some type of sensor such as a piezoelectric crystal or a microphone.

Aeolian sediment may be rolling or sliding along the ground (creep), bouncing in relatively short hops (saltation), or suspended for great distances (suspension) before returning to the ground. Most samplers are optimized to capture sediment transported by only one mode of transport. A recent survey of field samplers has been provided by Zobeck (2002).

Sampling efficiency refers to how well a sampler collects sediment compared to the actual amount of sediment in the wind stream. The sampling efficiency of a sampler must be known if estimates of true sediment transport are needed. An ideal sampler would have 100 per cent efficiency that does not change with wind speed or particle diameter. Unfortunately, most samplers have a range of efficiencies varying with wind speed and particle diameter. Several studies have described the efficiencies of saltation and suspended dust samplers (Shao *et al*., 1993; Goossens *et al*., 2000; Goossens and Offer, 2000).

Creep sediment samplers

Estimates of creep may be most economically made by simply burying a bottle with the opening flush with the soil surface as suggested by Bagnold (1941). A container with a round opening is necessary in field studies since the width of the opening will be the same from every direction. A more complicated but reliable creep/saltation sampler that orients into the wind has also been used successfully by Stout and Zobeck (1996). This near-surface passive creep/sampler collects material from three inlets at 0–3 mm, 3–9 mm, and 9–20 mm above the soil surface (Figure 6).

Saltation sediment samplers and sensors

Saltating sediment may be measured directly using sediment samplers or inferred using electronic sensors. Saltation sediment samplers are often called sand traps. The BSNE and MWAC samplers appear to be the most popular sediment samplers for field studies of saltation (Table I). Both types of passive samplers are easily mounted on poles to allow sampling at multiple heights. The BSNE is a wedge-shaped sampler with a 60 mesh screen on the top or sides to allow airflow out of the sampler to achieve near-isokinetic conditions (Fryrear, 1986). In most applications, a series of BSNE samplers are fixed to a pole and range in height from 0·15 to 1·0 m (Figure 1). A schematic of MWAC samplers mounted on a pole is shown in Figure 7. The BSNE sampler orifice

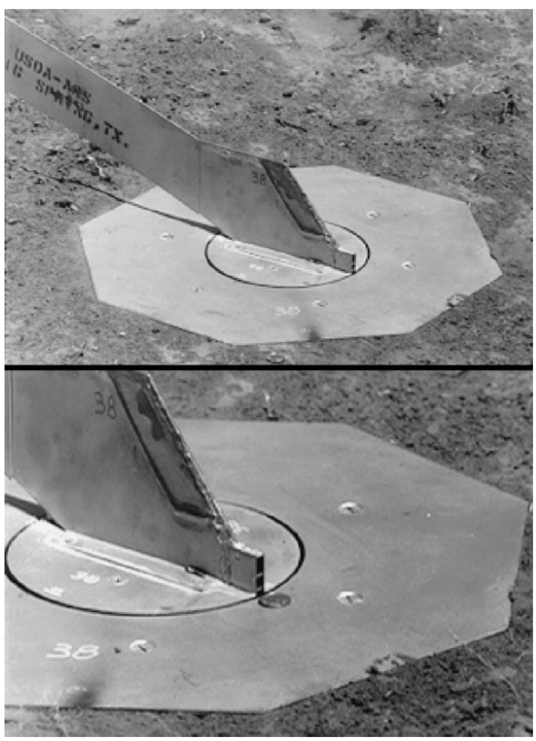

Figure 6. Near-surface creep/saltation sampler (reprinted from Zobeck (2002), p. 504, by courtesy of Marcel Dekker Inc.)

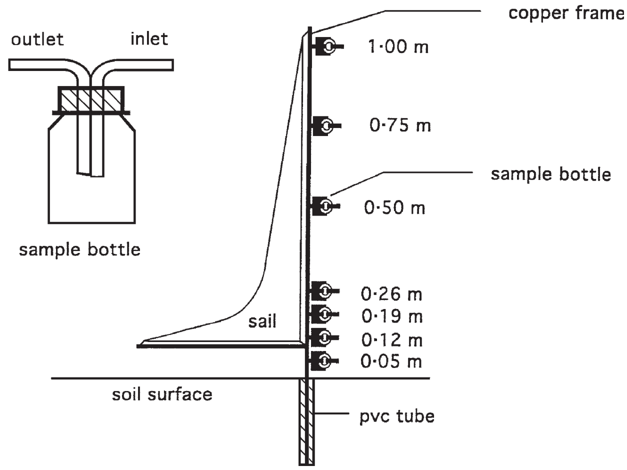

Figure 7. Modified Wilson and Cooke (MWAC) sediment catcher (pvc, polyvinyl chloride) (from Sterk and Raats (1996) with permission of the Soil Science Society of America)

is much larger (1000 mm^2 or 200 mm^2 for those designed for the near-surface measurements) than the MWAC (50 mm^2) and will collect larger samples. In detailed tests of five aeolian sand traps, Goossens *et al*. (2000) found that although both samplers had similar sampling efficiencies (70–135 per cent), the efficiency of the MWAC sampler was less influenced by ambient wind speed.

Although experimental saltating sediment samplers with automated, high frequency sampling capability have been described (Funk, 1995; Bauer and Namikas, 1998; Funk *et al*., in press), most studies use devices that are sampled manually after a storm. A few studies have manually sampled at intervals within a single storm (e.g. Stout and Zobeck, 1996).

Precise information on the initiation and duration of saltation has been gathered using commercially available electronic sensors that detect particle movement (Spaan and Van der Abeele, 1991; Stout and Zobeck, 1997). A SENSIT detects particles using a piezoelectric sensing element (Figure 2). A similar new particle sensing device called 'Safire', also utilizing a piezoelectric sensing element, has recently been described and tested (Baas, 2002). A Saltiphone is an acoustic sensor that detects particles striking a 200 mm^2 diameter microphone mounted in a tube with a wind vane (Figure 8). All electronic sensors detect particles at a frequency of 1 Hz or faster. The SENSIT has the advantage that it can be mounted very close to the soil surface and has an omnidirectional sensing element.

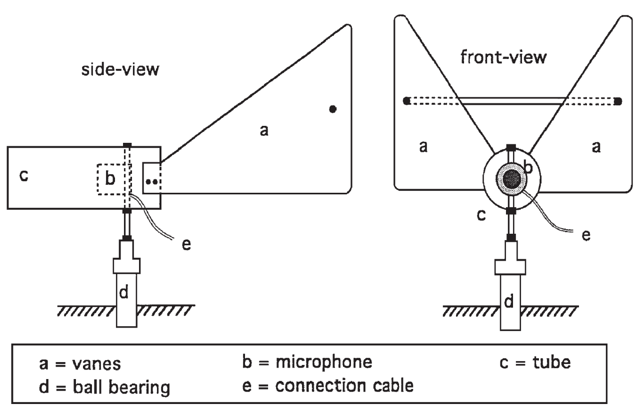

Figure 8. Saltiphone particle impact sensor (from Sterk *et al.* (1998) with permission of John Wiley & Sons, Ltd.)

Suspended sediment samplers

Passively sampling suspended sediment is more difficult than sampling saltating sediment because fine suspended particles (1) are easily carried by the wind stream and may not enter the sampler if it is not isokinetic, and (2) are not easily trapped by a screen or other physical barrier. Tests of several samplers at low wind speeds (1–5 m s^{-1}) using a silty loess that consisted of 95 per cent silt (2–63 μm) found a BSNE sampling efficiency of 40 per cent and a MWAC sampling efficiency of about 80 per cent for trapping suspended sediment (Goossens and Offer, 2000). However, sampling efficiency of the MWAC varied with wind speed. Sampling efficiencies at much higher velocities common in dust storms were not determined.

Studies addressing the emission of dust particles in suspension generally use active samplers that provide a suction using a pump of some type (e.g. Langham *et al*., 1938; Chepil and Woodruff, 1957; Gillette *et al*., 1974). For isokinetic sampling, the suction of the sampler is adjusted so the wind speed in the sampling orifice is the same as the ambient wind speed. Theoretically such a device is 100 per cent efficient. Dust concentration is usually determined by measuring the mass of dust deposited upon pre-weighed filters through which a known volume of air has passed. Nickling and Gillies (1993) describe a suspended sediment sampler with a 1·3 cm sampling orifice diameter that orients into the wind. Suction is provided by a high volume pump that is manually adjusted to match the ambient wind speed. A recent modification of this method was made by attaching the sediment sampling head to a DustTrak aerosol monitor described below (W. G. Nickling, personal communication; Figure 9).

Many current studies aim to quantify and characterize only a specific particle-size population, typically particles smaller than 10 μm (PM_{10}). Commercially manufactured high-volume (General Metal Works Inc.) and low-volume omni-directional samplers (AirMetrics, Inc.) were used in recent studies of emissions of PM_{10} from agricultural land (Stetler *et al*., 1994). Generally these devices are designed to measure dust concentration at low wind speeds (<10 m s^{-1}). Stetler *et al*. (1994) found that their instruments were accurate at wind speeds up to approximately 13·5 m s^{-1}. Modifications to a stacked filter unit were proposed by Cahill *et al*. (1996) to allow sampling up to 30 m s^{-1}. Nevertheless, most active suspension sampler efficiencies are not known for high wind

Figure 9. Suspended dust sampler (photo by W. G. Nickling)

speeds that occur during wind erosion events, specifically for mineral dust particles emitted during wind erosion of soil surfaces.

Although use of pre-weighed filters to determine dust mass gravimetrically is probably the easiest and cheapest method, dust mass can also be determined by other means. Mass can also be obtained by total elemental analyses of sediment on the filter by proton-induced X-ray emission (PIXE) (Johansson and Campbell, 1988) or X-ray fluorescent spectrometry (XRF) (Quisefit *et al*., 1994). The total mass of mineral aerosol is calculated by summing oxide forms of major elements. In the latter case, the total mass of sediment is underestimated because loss on ignition (LOI) is not taken into account. Due to the high sensitivity of these analytical methods (a few micrograms per filter) they are recommended to determine mass concentration over short periods of erosion such as those used to measure vertical flux (typically 20 min) (Rajot *et al*., 2003; Gomes *et al*., 2003a). Particle number concentration can also be obtained by direct counting of particles on filters by coupling microscopy and image analysis (Gillette *et al*., 1974). This method requires counting a large number of particles to ensure optimal particle sampling on the filters, especially if automated image analysis is used.

Commercially produced instruments are now available to indirectly measure suspended dust at rapid sampling rates (1 Hz). The DataRAM (MIE, Inc.), and DustTrak (TSI, Inc.) are aerosol monitors that measure aerosol concentration by light scattering and the GRIMM Environmental Dust Monitor (GRIMM Technologies, inc.) uses light scattering to measure particle concentration and size. The Tapered Element Oscillating Microbalance (TEOM) continuously measures mass of a filter directly during air filtration by means of a microbalance (Patashnick and Rupprecht, 1991). It is equipped with a classical low volume PM_{10} sampling inlet. While these instruments offer promise for detailed temporal measurements of dust during a storm, they also were designed for low wind conditions. Sampling efficiencies of these instruments for wind speeds occurring during a dust storm are not yet known but may be quite low. For example, a study of the sampling efficiency of the DustTrak shows the sampling efficiency varies considerable by particle size and wind speed (TSI, 2002). The sampling efficiency of 10 μm particles at a wind speed of 2·2 m s^{-1} was about 100 per cent while the sampling efficiency was about 40 per cent at a wind speed of 10 m s^{-1}. The sampling efficiency for 14 μm diameter particles was less than 10 per cent at a wind speed of 10 m s^{-1}. The study did not investigate wind velocities greater than 10 m s^{-1}.

Sampling for calculation of mass flux

Sampling the entire saltation layer or at multiple heights within the saltation layer is necessary to determine total horizontal sediment flux. For wind erosion at the scale of the field, the majority of mass flux will occur as creep and saltation material very close to the soil surface. For example, mass transport studies of a fine sandy loam soil showed that 50 per cent of the total mass transport occurred below a height of 1·7 cm (Stout and Zobeck, 1996). It is difficult to place a limit on the boundary between the zone dominated by saltation flow and that dominated by suspension flow. Stout and Zobeck (1996) showed that at a height of 70 cm, over 88 per cent of the sample collected by BSNE samplers had a diameter of less than 90 μm. Samplers rarely need to exceed 1 m in height for studies of saltation flux. Sampling at several heights, including the surface to measure creep, up to a height of 1 m will generally ensure capture of over 99 per cent of the creep/saltation sediment (see Stout and Zobeck, 1996, table 3). Estimates of mass transport may be biased when creep is not measured since creep can make up to 40 per cent of the transported mass.

Figure 10 illustrates the variation of aeolian sediment with height (Stout and Zobeck, 1996). In this example, eight saltation observations were made from 0·001 to 1·7 m above the soil surface using the saltation/creep and BSNE samplers. Several observations of aeolian sediment at different heights are needed to define the mathematical relationship of sediment flux with height. In this example, the saltation flux profile is very well described with a power-law equation. Detailed description of equations used to determine sediment flux is beyond the scope of this paper. Note the large sediment flux occurring very close to the soil surface. This example illustrates the importance of collecting samples near the soil surface. The lowest samplers at heights of 0·001, 0·006 and 0·015 m collected a large fraction of the total sediment in saltation. Not measuring these heights would result in a very different equation to describe the flux throughout the saltation zone. If only measurements of the total saltation transport are needed, a vertical slot sampler with a single vertical slot that spans the saltation zone, such as that described by Nickling and McKenna Neuman (1997) or a Bagnold catcher (Zobeck, 2002), might be a better choice.

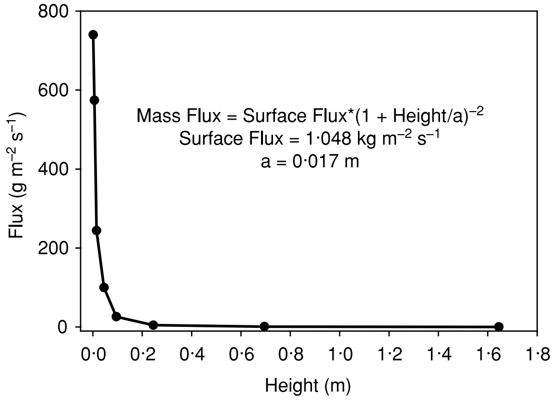

Figure 10. Vertical distribution of mass flux of an eroding fine sandy loam soil

Sampling the same particle sizes at two levels above the saltation layer is necessary to determine the vertical flux of dust. At least two heights of suspended dust sampling are necessary to calculate vertical dust flux using the gradient method and flux equation derived by Gillette (1977) (e.g. Nickling and Gillies 1993; Stetler and Saxton, 1996; Lopez *et al*., 1998; Rajot *et al*., 2003). One sampler must be placed above the saltation layer (about 1 m), and the other one a few metres higher depending on the fetch (typically from 3 to 10 m height). This method is based on the low sedimentation velocity of dust particles in comparison to the vertical velocity fluctuations during erosion. This property is assumed to be the same for small dust particles (diameter <20 μm) in typical wind erosion conditions, but not for larger ones (Gillette *et al*., 1972; Gillette, 1977). Particular care should be taken to ensure instruments used at both heights are calibrated to obtain the same results under the same experimental conditions. Small calibration errors among instruments may produce poor flux estimates.

Point or spatial sampling

The location and number of samplers used for field-scale studies will depend upon the objectives of the study and local meteorological and budgetary considerations. To measure saltation sediment transport across an area, the minimum number of sampler locations is two if the major erosive wind direction is relatively constant, one to measure the incoming flow and the second to measure the sediment leaving the site. The difference between the incoming and outgoing sediment is the eroded or deposited mass of the area lying between the two measurement locations. This approach was recently used by Bielders *et al*. (2000) to measure the effect of various on-farm management practices to reduce wind erosion in Niger. If wind direction varies from event to event, it is necessary to have sample locations along the study area boundary (Rajot *et al*., 2002). As described earlier in this paper, it is important in these studies to ensure that the field length used to compare among treatments is the same or has been scaled to the same distance among treatments. Additional samplers placed between these field border samplers can be used to determine the state of mass transport across the field.

Sampler location to determine the variability of sediment flux throughout the field is more complicated. Sterk and Stein (1997) applied geostatistical techniques to describe spatial variations of sediment flux throughout an experimental plot using 21 MWAC sediment catchers in a regular grid of 40 × 60 m. During one season four wind storms were observed, and the observed sediment mass fluxes showed large spatial variability. Despite the relatively large number of samplers for such a small plot size, 21 observations were not sufficient to estimate variograms for each storm (Webster and Oliver, 1992). Therefore the observations were normalized and pooled to allow geostatistical analyses. The obtained variogram was used to produce storm-based maps of sediment mass flux with kriging. These maps were used to determine accurate sediment mass balances for the plot.

Recently (G. Sterk, 2001, unpublished data), the same data and variogram were used to analyse the required number of saltation samplers for the 40 × 60 m plot. This was done with stochastic simulation. Possible realizations of sediment mass flux during a particular storm were generated using all 21 observations followed by simulations in which the number of observations was reduced each time. If ten observations, which were regularly divided over the plot, were used, the simulation results were statistically similar to those with all 21 observations. Hence, a minimum of ten catchers were required to obtain sufficient information on spatial variation within such a plot. Current wind erosion and dust transport models do not address this level of detail in spatial variability. Such detailed studies are needed for accurate mass budget estimates as well as model validation.

Other work on spatial sampling of wind erosion used nested sampling techniques (Chappell *et al*., 1996; Chappell, 1998; Chappell and Warren, 2003). A study of geostatistical methodology for estimating sediment transport, presented in this Special Issue (Chappell *et al*., this issue), tested nested, grid, and random sampling networks and recommended nested networks for estimation and mapping of sediment transport when few resources are available and especially for use over large areas.

In most cases, vertical dust flux is a function of horizontal saltation flux (Gillette, 1977; Shao *et al*., 1996; Alfaro and Gomes, 2001) and it is recommended that both horizontal saltation flux and vertical dust flux be measured at the same location to establish a relationship. Although the relationship of vertical and horizontal flux does not depend on fetch and can theoretically be obtained anywhere within the field, we further recommend installing suspension samplers at the place where the horizontal flux is at a maximum. In the case of variable wind direction, samplers should be placed in the centre of the field to maximize fetch from all directions.

Suspended dust samples also should be collected upwind of the field to establish the amount of dust from upwind sources. In general, although vertical suspension flux will have a strong relation to the local saltation flux, occasionally the source of some of the measured suspension flux in a study field is a distant upwind source where saltation activity is much higher. In addition, not all suspended dust may be caused by saltation activity. Recent studies, in a region of silty loess deposits where saltation does not appear to be a major factor, suggest suspended dust may also occur as a direct response to the wind (Kjelgaard *et al*., 2002).

Aeolian sediment sampling basic operating principles and strategies

Ideally, aeolian sediment samplers should be isokinetic with a high sampling efficiency. The sampling efficiency must be known to estimate the true sediment flux. Passive samplers are most commonly used to gather creep and saltation data and active samplers are most commonly used to collect suspended dust. Electronic impact sensors are recommended to estimate the onset and duration of erosion. Suspended dust samplers are commercially available but research is needed to determine their efficiency in high wind speeds common during wind erosion events.

Sampling the entire vertical column or at several heights, including the surface to measure creep, up to a height of 1 m is necessary to ensure capture of the creep/saltation sediment and determine horizontal sediment flux. Several observations of aeolian sediment at different heights are needed to define the mathematical relationship of sediment flux with height.

The location and number of samplers used for field-scale studies will depend upon the objectives of the study and the local meteorological conditions. To measure saltation sediment transport across an area, the minimum number of sampler locations is two if major erosive wind direction is relatively constant. If wind direction varies from event to event, it is necessary to have sample locations along the study area boundary. Variation of sediment flux within the field requires more complicated sampling patterns using random, nested, gridded or other networks.

METEOROLOGICAL MEASUREMENTS AND CALCULATIONS

Meteorological station instrumentation

Since wind erosion is an atmospheric process, careful measurements of a limited number of climatic variables are needed. The meteorological station should include an anemometer and wind vane, air temperature, solar radiation and relative humidity sensors, rain gauge, soil temperature and moisture sensor, and a data logger. All of the above instruments are available commercially and can be custom configured. These instruments are recommended to provide data used in erosion models (Table II). In many studies, meteorological instruments are mounted on 10 m towers. However, since most eroding fields are relatively flat with little standing vegetative cover, a 2 m tower is often adequate (Figure 11). Taller towers are recommended when studying tall standing vegetation.

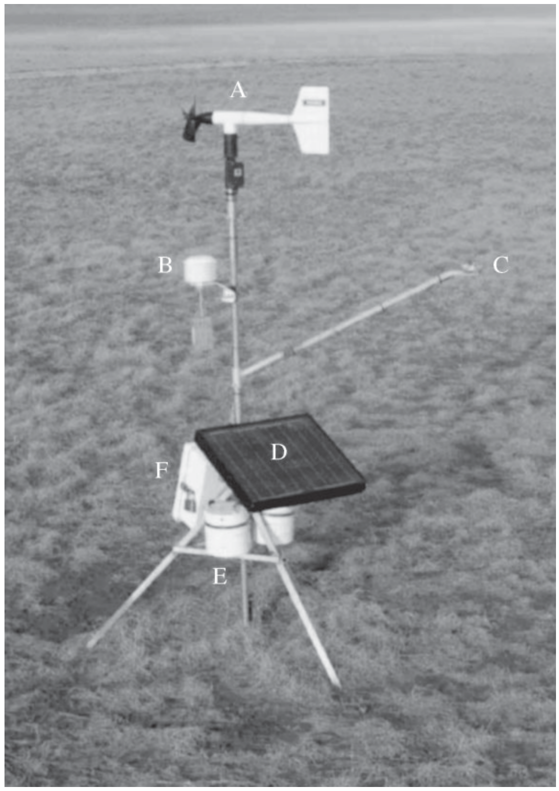

Figure 11. Meteorological tower, 2 m tall: A, combination anemometer and wind vane; B, relative and air temperature sensors; C, solar radiation sensor; D, solar panel; E, storage batteries; F, Data logger housing

Calculation of wind speed, friction velocity and aerodynamic roughness

Wind speed, friction velocity and aerodynamic roughness are the most important atmospheric parameters considered in most wind erosion studies. Since the standard height for measuring wind speed is 10 m, the 1/7-power-law profile (Simiu and Scanlan, 1978) can be used to convert data measured with smaller 2 m towers to a height of 10 m using the equation:

$$(u_2/u_1) = (z_2/z_1)^{1/7} \tag{5}$$

where u_1 and u_2 refer to the wind speed at heights z_1 and z_2, respectively.

In some cases where atmospheric stability parameters are desired, four or more anemometers are mounted on the meteorological tower along with concomitant air temperature sensors. High frequency measurements of temperature and wind speed allow the calculation of atmospheric stability parameters (Monin and Obukhov, 1954); however, enormous amounts of data are collected that are seldom needed for many wind erosion studies. Such information is more useful in studies concerning dust production (e.g. Lopez *et al*., 1998; Gomes *et al*., 2003b). During the strong winds of erosion events, the boundary layer near the surface is often statically neutral and corrections for stability are not necessary. However, the anemometers also allow calculation of the friction velocity (u_*) and aerodynamic roughness length (z_0) from the semi-logarithmic equation:

$$u(z) = \frac{u_*}{k}\ln\left(\frac{z}{z_0}\right) \tag{6}$$

where $u(z)$ (m s^{-1}) is the average wind speed at height z(m); u_*(m s^{-1}) is friction velocity; k is von Karman's constant (0·4), a dimensionless number; and z_0 (m) is the aerodynamic roughness height. The latter corresponds

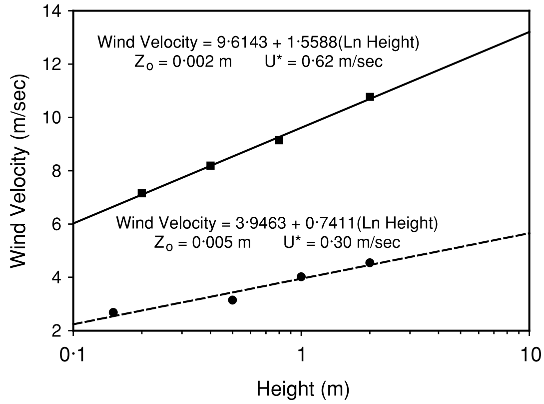

Figure 12. Wind velocity as a function of height for two different soil surfaces

to the theoretical height at which wind speed near the surface falls to zero and depends on the roughness of the ground surface. The friction velocity is a measure of shear stress at the surface, and is used as the driving variable for windblown sediment transport in most wind erosion models. The aerodynamic roughness parameter z_0 is useful to determine soil erodibility conditions. In general, z_0 increases with surface roughness, leading to decreased soil erodibility. Equation 6 applies to the surface portion of the planetary boundary layer where Coriolis effects can be ignored and the momentum flux may be considered constant, independent of height. Furthermore, it does not consider any viscous sublayer that may exist over a smooth surface or the canopy or roughness sublayer in which the flow is very likely to be disturbed by individual roughness elements (Arya, 1988).

The determination of parameters u_* and z_0 is facilitated by rearranging terms to create the linear equation form:

$$u(z) = \left[\frac{u_*}{k}\right]\ln(z) - \left[\frac{u_*}{k}\right]\ln(z_0) \tag{7}$$

A plot of average wind speed versus height as shown in Figure 12 illustrates wind speed data for two different surfaces plotted using Equation 7. The parameters are determined from the slope and intercept of the line. Multiplication of the slope (u_*/k) by 0·4 defines the friction velocity, u_*. The aerodynamic roughness is found as the anti-log of the quotient of the intercept and slope.

Soil surface conditions that affect the wind profile may vary spatially and should be considered in field-scale studies. A single stationary meteorological tower may not adequately describe the spatial variability of u_* and z_0. Information on the spatial variability of these parameters is easily measured using a portable 2 m tower that can be moved around the field (Figure 13). The tower shown in Figure 13 includes a wind vane and four anemometers. Measurements of the wind profile are needed after each significant change in aerodynamic roughness, such as that produced by a growing crop.

Sampling interval

The sampling interval for the meteorological instruments varies depending upon the purpose of the study. A one-hour sampling interval is adequate to document seasonal variation in erosion. For studies at the time scale of one erosion event, the sampling interval should not exceed 5 min. Even shorter sampling intervals are necessary to arrive at the true value of erosion threshold wind speed. Calculated threshold values decrease as wind speed averaging time increases (Stout, 1998). Stout (1998) found that threshold estimates based on a 10 s averaging time were about 18 per cent less than estimates based on a 1 s averaging time. The recommended sampling interval to estimate erosion threshold wind speed is 1 s.

Some studies use a combination of sampling intervals. In the Columbia Plateau Study (Stetler *et al*., 1994; Saxton, 1995), all operations were controlled by a data logger which started an electrical power generator and PM_{10} sampling devices when the wind velocity at 2 m height exceeded a 15 min average of 6·7 m s^{-1}. The system

Figure 13. Anemometer tower, 2 m tall

shut down after the 15 min average wind velocity fell below the threshold. This technique reduces the amount of computer memory needed but may introduce some errors since part of a storm may be missed and storm duration miscalculated when using a 1 h average. In addition, it is often more difficult to process the data when variable time intervals are used.

Meteorological basic operating principles and strategies

Careful measurements of a limited number of climatic variables are needed. Meteorological measurements should include wind speed and direction, air temperature, solar radiation, relative humidity, rain amount, soil temperature and moisture. Wind speed, friction velocity and aerodynamic roughness are the most important atmospheric parameters considered in most wind erosion studies. Measurements of the wind profile are needed after each significant change in aerodynamic roughness, such as that produced by a growing crop.

CONCLUSIONS

Wind erosion is a natural process that has only been studied intensively for about the last half century. Measurements of wind erosion at the field scale can be done in many ways, using different types of sampler, field sizes, experimental layouts and analysis techniques. However, although the methods and instruments are numerous, the selection of the instruments and methods employed are governed by relatively few operating principles. In this paper we have outlined important principles to consider in conducting field-scale wind erosion studies and proposed strategies of field data collection for use in model validation and development. Methods and instrumentation used are often a matter of preference or simply based on the availability of equipment and budget constraints. However, adherence to the principles outlined will facilitate future use of these data by others.

For validation of dust emission models, there is a need for coupled horizontal and vertical erosion flux measurements (Alfaro *et al*., 2002; Saxton *et al*., 2000). This requires performing vertical dust flux measurements in field studies. Such measurements would become easier by the development of new automated dust monitoring instruments. The new instruments need to measure suspended dust concentration with high efficiencies at wind speeds of at least 15 m s^{-1}. Coupled measurements also necessitate measurements of horizontal flux resolved at the same time periods as measurements of vertical flux. Efforts are needed to produce an automated device to measure saltation flux with high sampling frequency. Some success has been possible in calibrating saltation impact sensors such as SENSIT (Gillette *et al*., 1997) and Saltiphone (Rajot *et al*., 2003).

To improve comparison of field-scale study results, we suggest that studies include a minimum of important details on the experimental conditions. The most important items to be listed include:

(1) *field data*: location and dimensions, direction of tillage, occurrence and description of any upwind obstructions;
(2) *sediment sampler data*: number, type, placement, efficiency, sampling frequency, fetch distance, time of sampling;
(3) *soil surface data*: soil type and classification, texture, soil moisture, organic matter and calcium carbonate content, random and oriented roughness, dry aggregate size distribution, presence of crust and estimate of stability and loose erodible material, soil cover type and amount;
(4) *meteorological data*: wind speed and direction with averaging times during storms, aerodynamic roughness, friction velocity, duration of storms, antecedent rainfall, relative humidity, solar radiation, air temperature, and wind direction variability during storm sampling.

Careful consideration of the climatic, sediment, and soil surface characteristics observed in future field-scale wind erosion studies will ensure maximum use of the data collected.

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